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Anna Stepanovna Tarabanova, a permanent scientific editor of the Journal Stratigraphy. Geological Correlation turned 70 years old on April 17, 2002. For 45-year work in the Publishing House Nauka, Anna Stepanovna proved to be an industrious and responsible associate offering a professional approach to her work, as well as responsive and well-meaning traits.

Dear Anna Stepanovna! The Editorial Board of the Journal Stratigraphy and Geological Correlation congratulates you with a renowned jubilee. We wish you with all our hearts much health and well-being at home and at work, the joy of life and good spirits.

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Geochronological Scales and the Astronomic Time (State of the Art)

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Abstract—A problem of modifying the International stratigraphic scale is discussed, and the present state of the Phanerozoic geochronological scale is analyzed. It is suggested that both scales reflect a combined response of the Earth subsystems (tectonosphere, biosphere, hydrosphere, and others) to powerful cosmic impacts on our planet, an element of the Solar System. The proposed ideas are supported by a theoretical model allowing to relate epochs of mass extinction of living organisms to periods of bombarding the Solar System by galactic comets and to moments of big asteroid impacts on the Earth. The proposed geological time scale that is calibrated with proper allowance for distribution of gravitational potential in the Galaxy is comparable in accuracy with the Phanerozoic geochronological scale.

Key words: geochronological and stratigraphic scales, cosmic impacts, mass extinction events, jet streams of the Galaxy, galactic comets, asteroids.

INTRODUCTION

A century ago, in 1900, the VIII Session of the International Geological Congress adopted the first version of the International stratigraphic scale (ISS). Since that time the ISS has been a subject of continuous modification. The stratigraphic scale is commonly considered (Leonov, 1973) as a reflection of natural stages of the geological history of the Earth.

As time passed, the non-uniform character of the ISS became clear. The heterogeneity is evident from the fact that its Precambrian, Phanerozoic, and Quaternary parts differ in the time range of units and in the leading method of subdivision. For instance, stratigraphic units of the Precambrian span hundreds of million years and mostly correspond to tectono-magmatic cycles. The Phanerozoic stratigraphic units spanning millions of years are distinguished based on organic evolution. A leading method of subdividing the Quaternary Period is climatostratigraphy.

Based on fossil records, the Phanerozoic stratigraphic scale is unique, having the highest resolution. At present, it is divided into stages and even zones and serves as a test scale for different methods of subdivision and correlation of deposits. These methods revealed the non-uniformity of systems, series and stages of the ISS.

The recent rapid development of stratigraphic investigations in vast regions of continents and oceans, and the relevant improvement of the ISS opened a possibility to elaborate a standard stratigraphic scale of global significance. At present, there are at least 15 variants of the scale, which are based on different approaches and methods, and some of them quite dissimilar with respect to boundary positions and time ranges of stratigraphic units. Therefore, to avoid errors in dating and, in turn, mistaken correlations, the boundary ages are usually given with reference to the scale used.

CHRONOMETRIC, STRATIGRAPHIC, AND GEOCHRONOLOGIC SCALES

The accurately established succession of stratigraphic units reflects an order of their formation with time. However, an important disadvantage of these stratigraphic scale is an ambiguity of boundaries and scopes of subdivisions, which are hardly recognizable beyond the stratotype region. To remedy the uncertainty, some biostratigraphic, lithologic and other features of individual stratotype sections, which are thought to reflect global events, such as alteration of "warm" and "cold" climatic periods, noticeable sea level oscillations, geomagnetic reversals, and so on, are taken into account. The impact moments of big asteroids and comets falling on the Earth can be of similar significance. If the chronometrical succession of different geological events would be depicted against a single stratigraphic background, this could essentially improve the scale.

The biostratigraphic method is of primary importance for practice. This method permits the detailed subdivision and most reliable correlation of deposits in

different regions, and elaboration of stratigraphic schemes of the regional and global significance (Gladenkov, 1998).

The subdivision and correlation are most successful when the stratigraphic scale based on relative ages of units is used together with the geochronological one. The geological practice showed long ago that the geologic time scale, which is called geochronological, is dual in essence (Menner, 1979; Meyen, 1989). It virtually unites stratigraphic and chronometric scales. Although being permanently modified, the stratigraphic scale is relatively stable, whereas chronometrical ages of units and their boundaries experience revision from time to time.

THE PRESENT STATE OF PHANEROZOIC GEOCHRONOLOGICAL SCALE

A version of the geochronological scale for the Phanerozoic elaborated by the London Geological Society in 1964 and edited by W. Harland was first and very complete. The scale appeared to be very adequate (Schindewolf, 1970; Sokolov, 1971) and was partly incorporated into all subsequent scales. It was calibrated based on age dates for 337 objects supplemented afterwards by 29 objects (Harland *et al.*, 1971).

Using new and revised isotopic dates recalculated in accord with newly estimated decay constants, Harland *et al.* (1982) proposed a new variant of the scale. Another version of geochronological scale compiled by Odin (1982) was published in the same year. It was based on isotopic dates of 251 objects, among which only seven dates were sufficiently reliable (Rublev, 1996).

The principal difference between two versions is the dissimilar level of the Phanerozoic lower boundary that is accepted to be 590 Ma old by Harland with colleagues and 540 Ma old by Odin. In addition, dates for the Cretaceous and Silurian lower boundaries differ by 14 and 20 m.y., respectively.

Other versions of the geochronological scale, which appeared somewhat later, are again incompatible with respect to time ranges of their subdivisions. Snelling (1985) published a very popular scale, which is commonly considered as a compilation of scales of 1982. In opinion of many scientists, the scale of Palmer (1983) is of the same type. In the new scale of Harland *et al.* (1990), 163 of 670 isotopic dates are borrowed from their previous variant (Harland *et al.*, 1982) and 278 are from the Odin's scale (Rublev, 1996).

Isotopic ages obtained recently led to the revision of dates for some units. For instance, the Cambrian lower boundary that was universally accepted at the level of 570 Ma, is risen now to 540–545 Ma. The last result was used by Odin (1994) in a new variant of the Phanerozoic geochronological scale. The latter was followed by the new Cenozoic time scale of Berggren *et al.* (1995) and by the Phanerozoic time scale of Gradstein and Ogg (1996). Finally, the modified International stratigraphic chart (Table 1) was proposed in 1998.

Although all the geochronological scales mentioned above are based on combination of biostratigraphic and isotopic-geochronological data, they show discrepancies between boundaries of some subdivisions. Levels of some boundaries did not change during the last 20 years, whereas others were revised many times (Barenbaum and Yasamanov, 1999). Leaving aside some boundaries, which are conventionally displaced, recent geochronological scales include, in addition to those stratigraphic and chronometric levels, which are unambiguously defined, many others that are still controversial.

These discrepancies and parallel existence of several scales likely point to the absence of well-defined criteria enabling subdivision of stratigraphic units into smaller ones, particularly in the substages and zones. Universal principles of elaboration of the standard geological time scale for the whole Earth are also unclear (Simakov, 1999).

A NEW APPROACH TO ELABORATION OF THE GEOLOGICAL TIME SCALE

Ten years ago, the connection between cyclic occurrence of geological phenomena on the Earth and some processes in the Galaxy could be only hypothesized more or less successfully. There were no grounds to believe that astronomical sciences may significantly contribute to the solution of geological problems, including those of stratigraphy.

The situation radically changed when the hypothesis of J. Jeans on substance flows out of centers of the spiral galaxies had been used for explanation of geological events (Barenbaum, 1991). This hypothesis served as a base for a new system of notions developed recently and offering an opportunity to relate some processes on the Earth and in the Solar System to this phenomenon.

An important component of the system is a model that enables creation of "a galactic scale" of geological time (Barenbaum and Yasamanov, 1995, 1999) based mostly on astronomical observations and calculations. The model elucidates possible causes of global geological and biological catastrophes that occurred periodically and left, in faunal assemblages and lithological complexes, their signatures, which are usually considered as intrinsic boundaries of stratigraphic and geochronological scales. The model also offers a simple explanation to the principle of the existing scales subdivision into periods (systems), epochs (series), ages (stages), and so on (Barenbaum, 1994).

The main idea is that all global geological cycles having periods of ~10^6 to 10^9 years correspond to both the separate stages of the Earth evolution and some powerful cosmic events, which quasiperiodically occur in the Solar System with a great impact on terrestrial

Table 1. Ages of Phanerozoic stratigraphic boundaries according to different scales

System	Series	*Geokhronologicheskaya tablitsa* (1978)	Harland *et al.* (1982)	Palmer (1983)	Snelling (1985)	Harland *et al.* (1990)	Gradstein and Ogg (1996)	International stratigraphic chart (1998)
Q		1.8	2.0	1.64		1.64	1.8	
N	N2		5.1	5.2		5.2	5.3	5.3
	N1	23 ± 1	24.6	23.3		23.3	23.8	23 ± 0.5
P	P3		38.0	35.4		35.4	33.7	33.7 ± 0.5
	P2		54.9	56.5		56.5	54.8	53 ± 1
	P1	65 ± 3	65	65		65	65.0	65 ± 0.5
K	K2		97.5	97	95	97	98.9	96 ± 2
	K1	135 ± 10	144	145.6	135	145.6	142.0	135 ± 5
J	J3		163	157	152	157	159.4	154 ± 5
	J2		188	178	180	178	180.1	175
	J1	190 ± 5	213	208	205	208	205.7	203 ± 3
T	T3		231	235	230	235	227.4	230 ± 6
	T2		243	241	240	241	241.7	240 ± 5
	T1	230 ± 15	248	245	250	245	248.2	250 ± 5
P	P2		258	256	270	256	256.0	258 ± 5
	P1	295 ± 15	286	290	290	290	290.0	295 ± 5
C	C3			303	300	305		
	C2		320	323		323	323.0	315 ± 5
	C1	350 ± 15	360	362	355	362.5	354.0	355 ± 5
D	D3		374	377	375	377	370.0	375 ± 5
	D2		387	386	390	386	380.0	385 ± 8
	D1	405 ± 10	408	408.5	405	408.5	417.0	408 ± 5
S	S2			424			423.0	
	S1	435 ± 10	438	439	435	439	443.0	435 ± 5
O	O3		458	464	455	464	458.0	
	O2		468	476	460	476	470.0	
	O1	480 ± 15	505	510	510	510	495.0	
Є	Є3		523	517		517	505.0	
	Є2		540	536		536	518.0	
	Є1	570 ± 20	590	570	570	570	545.0	540

* Not all the scales mentioned in the article are presented.

processes. Therefore, present stratigraphic and geochronological scales with their hierarchy of variably long cycles superimposed one on the other are, in essence, the empirical classifications of transformations in the Earth subsystems (tectonosphere, hydrosphere, atmosphere, biosphere, and others) after powerful cosmic impacts on our and other plants of the Solar System. Throughout the history of the Earth, the main cosmic factor was bombarding of the Earth surface by big cosmic bodies: asteroids and galactic comets.

Interpretation of major geochronological boundaries, which correspond to cosmic impacts of supposed energy, is given in Table 2.

Geological data depicting four significant events in the Earth history (time marks at 3.6, 2.6, 1.65, and ~1.05 Ga) can be interpreted (Barenbaum and Yasamanov, 1995) as consequences of the Sun flight through the dense star clouds. The interactions between the Sun and stars caused "spasmodic" changes in the solar orbit parameters. These changes, in turn, greatly affected all processes in the Solar System. The cosmic events coincided with the most significant geological and biotic catastrophes in the Earth history, which defined records accepted for boundaries of major stratigraphic and geochronological units (eonothems and eons, Table 2).

Table 2. Chronological chart of the cosmic events

Cosmic event (physical cause)	Unit	Time (t) or period (T) of events	Energy of impacts, J
Sun flight across dense star clouds, abundant asteroids falling on the Earth	Eon, eonothem	t = 3.6, 2.6, 1.65, 1.05 Ga	~10^{31} ~10^{28}–10^{29}
Bombardment of the Solar System by galactic comets			
"Strong"	Era, erathem	T = 250 m.y.	~10^{29}
"Median"	Period, system	$T \approx$ 20–40 m.y.	~10^{27}–10^{28}
"Weak"	Epoch, series	T = 19–37 m.y.	~10^{26}–10^{27}
Collision of the Earth with solitary asteroids			
diameter more than 3.5 km	Age, stage	$T \approx$ 2.9 m.y.	~10^{22}–10^{23}
diameter less than 3.5 km	Zone	$T \approx$ ~0.01–1.0 m.y.	<10^{22}

Two main mechanisms of affecting the Earth by star clouds are possible. In one case, a spasmodic turn of the Sun orbit at about several angular degrees (Barenbaum, 1991) gives the Earth the powerful energetic impulse of ~10^{31} J. This energy was mostly spent on changing the Earth orbit around the Sun and partly dispersed in the Earth interior as heat.

The other case is bombarding of the Earth surface by big asteroids. When the Sun abruptly changes its galactic orbit, abundant asteroids penetrate into the interplanetary space from the Asteroid Belt, and planets "capture" them during the periods of ~10^7–10^8 years. Asteroids fell on the Earth in abundance after the events of 3.6 and 2.6 Ga, less frequently after 1.65 Ga and still lesser after 1.05 Ga. On the Earth, the cosmic material was completely dispersed and mixed with terrestrial matter to cause extraordinary weathering processes and ore formation. The intensive weathering may also be a response to a great amount of free oxygen released in the atmosphere during the impact-induced evaporation (Markova *et al.*, 1986) of surface rocks.

In the periods between the Sun interaction with stars, the most powerful cosmic impacts on the Earth were produced by galactic comet showers. The comet bombarding took place when the Sun crossed the galactic jet streams with intervals of 19–37 Ma during the Phanerozoic time. The comet bombarding was most intense (boundaries of eras and erathems), when the Sun orbit crossed the spiral arms of the Galaxy (Fig. 1). Episodes of less intense bombarding are imprinted as boundaries of periods (systems) and minor ones as boundaries of epochs (series).

When the Solar System entered the jet streams (during ~1–5 m.y.), up to several hundreds of thousands of galactic comets in total may fall on our plant (Barenbaum, 1990). Icy cores of comets become destroyed in the Earth atmosphere (Grigoryan, 1974; Adushkin and Nemchinov, 1994). Their energy was spent to form a powerful impact wave that can produce enormous destruction and cause mass extinction of living organisms on the Earth surface, but it cannot create a big crater.

All known Earth craters of impact origin are the results of asteroid falling. Asteroids fall independently on galactic comets with different periodicity. The "comet showers" of different intensity are relatively rare, whereas big solitary asteroids fall much more frequently. Nearly 200 events of the Phanerozoic time can be interpreted (Barenbaum, 1994) as the impact results of solitary asteroids more than 3.5 km in diameter. The impacts were accidental, one per 2.9 m.y. in average. About 45% of these events are recorded as age (stage) boundaries. Other impacts (55%) may correspond to boundaries of finer geochronological units (chronozones), but being coincident with the comet bombarding, they can hardly be recognized.

The falling asteroids with critical diameter of 3.5 km have the energy of ~10^{22}–10^{23} J. The impacts of greater energy probably had global consequences that lasted for a long time. The more frequent impacts of a lower energy, which are recorded in a limited number of regions, were of a regional or local significance.

Let us discuss now systems and series (periods and epochs) of the Phanerozoic stratigraphic scale in terms of the galactic model.

GALACTIC MODEL

The basic postulate of the model is that the Galaxy, i.e., the evolving star system, is in a quasistationary state that is maintained by continuing destruction of stars in its center with subsequent ejection of gas and dust products in a form of jet streams. Gas and dust, dispersed throughout the galactic plane, condense to form the gaseous–dusty clouds, comets, and stars. The Sun orbiting the Galaxy center along a slowly changing elliptical path periodically crosses the jet streams and galactic spiral arms. At these moments, planets, the Earth included, experience impacts of bodies moving inside the jet streams, mostly of comets.

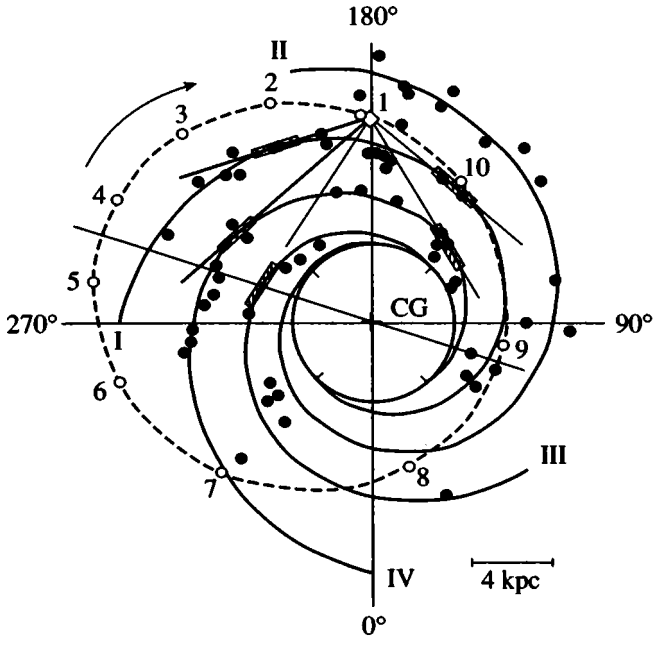

Fig. 1. The calculated Sun orbit (dashed line) in the spiral arms of the Galaxy: (CG) center of the Galaxy.

Numbered points of orbit indicate moments, when Sun is inside the jet streams of the Galaxy. Roman numbers denote spiral arms of the galactic plane, which extend through giant molecular clouds (black circles) and areas of high gas emission (hatched rectangles). Arrow shows direction of the Sun orbital motion and rotation of the spiral arms around the Galaxy center.

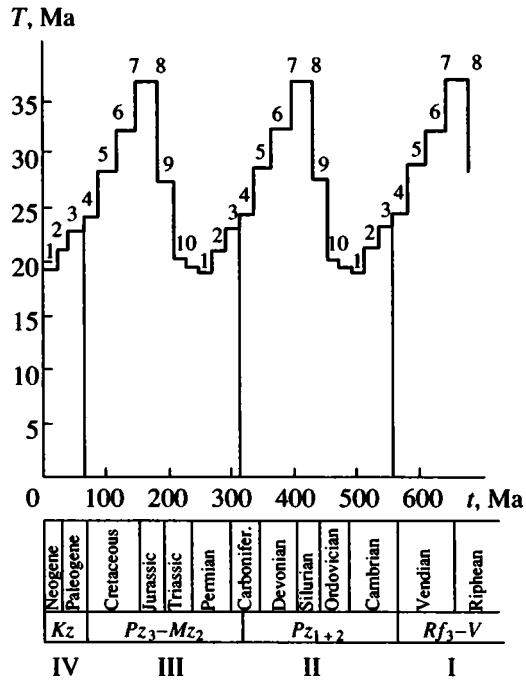

Fig. 2. Time intervals separating periods the Sun residence inside the jet streams of the Galaxy during the last 700 m.y. Presumable megacycles of the Sun rotation around the Galaxy center, which correspond to the latest Proterozoic (I), two Phanerozoic (II and III) and early Cenozoic (IV) periods. Numbers from 1 to 10) denote the Sun position in the orbit (Fig. 1). Universally accepted geochronological periods are shown below the diagram.

Connection of global geological processes with galactic events is established using a physical model to calculate the Sun orbit and to figure out the moments, when the Sun crossed the galactic jet streams during the last 700 m.y. The results of astronomical and geological observations were taken into account. Taken from astronomy were data on the Sun remoteness from the Galaxy center and spiral arms, and the tangential and radial components of the Sun velocity. The accepted levels of chronostratigraphic boundaries are those from the Geochronological Chart (*Geokhronologicheskaya* ..., 1978).

The calculations showed (Barenbaum, 1991) that even the simple Keplerian model was adequate, showing a perfect correspondence between the calculated data and ages of boundaries of virtually all geochronological periods and epochs. According to this model (Fig. 1), the Sun orbit projected onto the galactic plane represents an ellipsis having eccentricity of 0.36 and the major and minor axes of 10.2 and 9.5 kpc long, respectively. The orbital period of the Sun is 250 m.y. Fig. 1 shows 10 "peculiar" orbital points (numbered) where the Sun was inside the streams of galactic comets.

The calculated time intervals between moments of the Solar System bombardment comets, which correspond to time ranges of chronostratigraphic units, are shown in Fig. 2. Because of the uneven orbital motion of the Sun, the orbital period is modulated to be 250 m.y. The cycles show a certain trend of changing durations of stratigraphic units. If we position the cycle boundaries as shown in Fig. 2, the time ranges of units are less than 25 m.y. in the first half of the cycle and more than that in the second half. They are minimal (19 m.y.) for the Permian and Cambrian and maximal (37 m.y.) for the Jurassic, Silurian, and Late Riphean. These variations were characteristic not only of the Phanerozoic but also of the entire Late Proterozoic (Barenbaum and Yasamanov, 1995).

In accord with their periodicity, the above subdivisions can be united into larger units, such as eras (erathems), which are similarly variable in duration. It was expected that the time range of an era should correspond to the galactic year. However, it appeared that the commonly accepted time ranges of eras are different. The Paleozoic and Cenozoic eras (Point 4 in Fig. 1) begin with subdivisions, which differ in duration from those of the Mesozoic era (Point 10 in Fig. 1). If all the cycles would begin from Point 4, the boundary of the Mesozoic Era should be coincident with the Frasnian–Famennian boundary of the mid-Carboniferous, according to the galactic model.

The careful comparison of calculations with some widely recognized geochronological scales published before 1993 suggests that the above discrepancy is caused by a slow turn of the Sun orbit toward the direction of the Sun movement in the Galaxy (Barenbaum and Yasamanov, 1999). However, a simplified character of the model and ambiguous dating of stratigraphic boundaries by different authors exclude a possibility to solve this problem and to estimate the turn rate. In this work, the moments when the Sun crossed the jet streams and spiral arms of the Galaxy are calculated taking into account that turn of the solar orbit.

Analyzing new modeling results, we paid a keen attention to events of the biota mass extinction (ME). The last events seem to be more sensitive indicators of the comet bombarding epochs, than the activation periods of tectonic, magmatic, volcanic, climatic, and other geological processes. In addition, the recent dating of the Phanerozoic ME epochs is sufficiently precise for our purposes.

We are aware of numerous hypotheses explaining the particular biotic crises by different terrestrial (abiotic and biotic) and extraterrestrial causes (Alekseev, 1998). There are opinions (McLaren, 1983; Alekseev, 1993) that the ME events occurred under successive and combined influence of many factors of endogenic, exogenic, and cosmogenic origin. However, only one or two factors are always those that trigger the event. At present, the increasing number of researchers believes that the cosmic impacts are among the latter.

Bearing in mind all other hypotheses, we would like to consider here the extinction events in detail and to interpret them in terms of the galactic model.

THE MODIFIED GALACTIC MODEL

The galactic model was analyzed using the database of Sepkoski (1995) on the Phanerozoic marine biota extinctions. He summarized data on 34625 extinct genera of marine organisms from 1040 published and 10 unpublished works of his colleagues. The number of extinction events was calculated per each of 84 time units having mean duration of 6.4 m.y., which roughly correspond to ages of the Geologic Time Scale (Harland *et al.*, 1990).

Data summarized by Sepkoski show that extinction of organisms of lower taxonomic levels (species, genera, families) was not monotonous during the Phanerozoic. Against the slightly variable background of continuous extinction, there were separate peaks of "mass extinctions" (Fig. 3a). Some events that are called the "great" ME occurred at the Cretaceous–Paleogene, Triassic–Jurassic, Permian–Triassic, and Ordovician–Silurian boundaries (Veimarn *et al.*, 1998). In addition to these four events, there was one more great extinction (Alekseev, 1998) that lasted between the Frasnian–Famennian and Tournaisian–Visean boundary periods.

Table 3 illustrates how the major Phanerozoic ME events (Sepkoski, 1995; Alekseev, 1998) correlate with the moments when the Sun was inside the galactic jet streams, which are calculated on the basis of the Keplerian model (Barenbaum, 1991). The ME events are assigned to the geochronological scale of Gradstein and Ogg (1996).

As for data of Table 3, we should note that the moments of the Sun position in "centers" of the jet streams were accepted as basic ones. As marginal zones of a jet stream have certain width, the Sun enters it approximately 1–2 m.y. before reaching the center and leaves during the same time after crossing the latter. During this time interval the intensity of the comet bombarding may be extremely changeable. Some bombarding periods could be much longer, if we take into account that the Sun crossed both the jet streams and the considerably wider spiral arms in the same period.

Geological ages of the ME events are somewhat uncertain. These were quite long events that lasted 1–6 or even 10–15 m.y. (Alekseev, 1989). However, traditionally they are attributed either to a single age or to the boundary period between ages.

As was already mentioned (Table 2), the age boundaries mark the impacts on the Earth of critical-sized asteroids. In the Phanerozoic, the asteroids fell accidentally, each 2.9 m.y. in average. Therefore, in the intervals of several million years, when the Solar System occurred inside the galactic comet stream, one or several big asteroids might or might not fall on the Earth. As a result, the extinction of living organisms was specific in each epoch.

According to the classification of Alekseev (1998), the ME events without strictly determinable stratigraphic position are referred to "phantoms." In our model, phantoms are considered as results of the galactic comet falling. The examples are the Stephanian ME that lasted through the second half of the Kasimovian (303.0–296.5 Ma) until the Gzhelian inclusive (296.5–290.0 Ma) and another event that began in the Kimmeridgian (154.1–150.7 Ma), continued in the Tithonian (150.7–142.2 Ma) and terminated in the Berriasian (142.2–137 Ma). The terminal moments of both mark the Carboniferous–Permian and the Jurassic–Cretaceous boundaries, respectively (Table 1).

During the most ME epochs, at least one big asteroid fell on the Earth as a result of the comet bombardment. The impact moment was characterized by particularly significant extinction that lasted the whole epoch. If the asteroid fell at the beginning of the comet bombardment, it could cause the biotic crisis as it has been suggested (Alekseev, 1998).

As a rule however, asteroids fell after initiation of the comet bombardment, as exemplified by the well-established asteroid impact during the Cretaceous–Paleogene ME event (Alvarez *et al.*, 1980; Nazarov *et al.*, 1988; Veimarn *et al.*, 1998). The corresponding decrease in taxonomic diversity of marine organisms

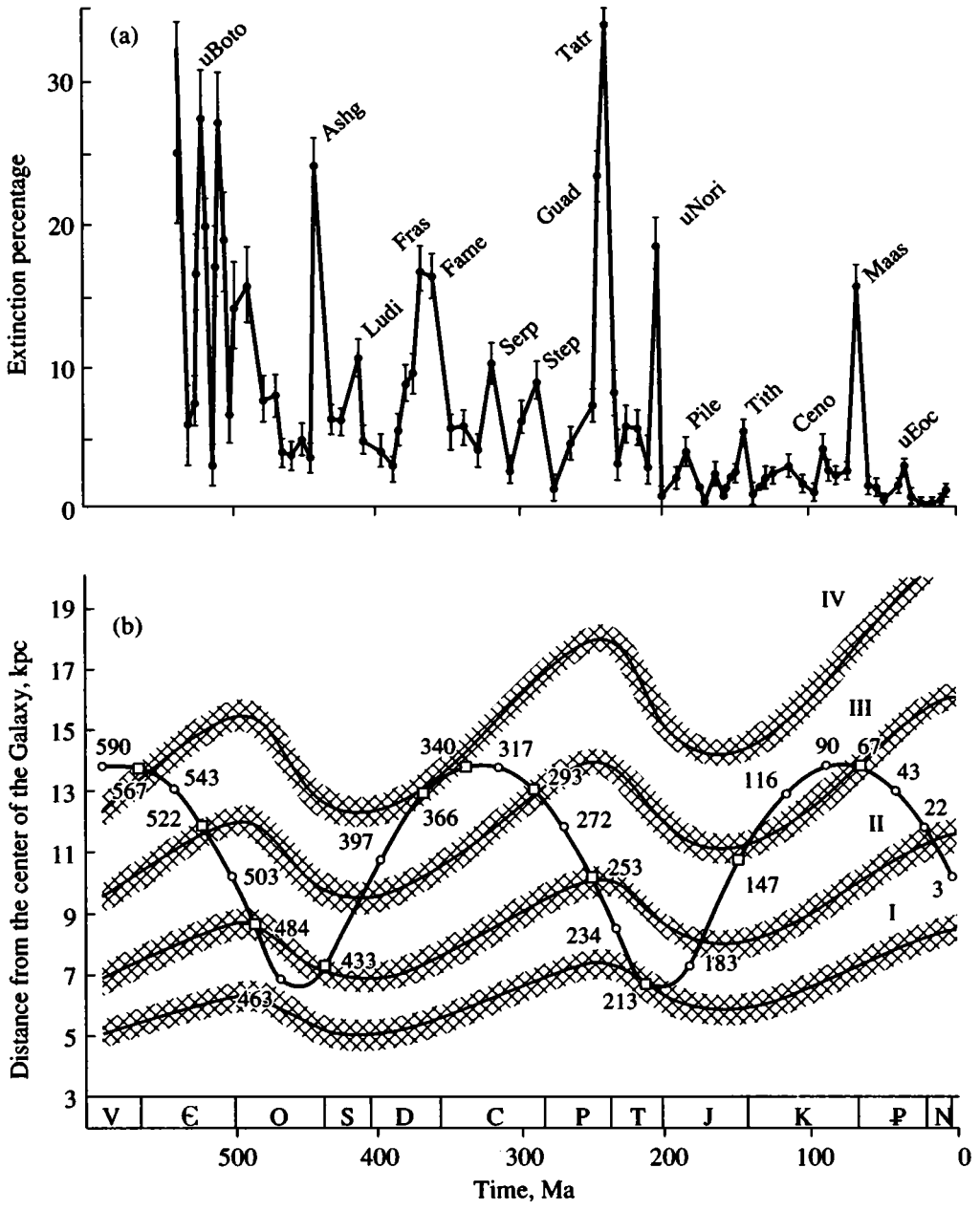

Fig. 3. Secular variation curve (a) of Phanerozoic marine organisms (families) according to Sepkoski and position of solar orbit relative to galactic center and four spiral arms (b). Roman numbers denote spiral arms of the Galaxy (see Fig. 1), which are conventionally considered as 1 kpc wide. Points on the Sun orbit (solid line) denote the time moments, when the Sun entered the jet streams (calculated with allowance for the solar orbit rotation). Moments when the Sun was simultaneously inside the jet streams and the spiral arms are indicated by squares.

began at the early–late Maastrichtian boundary (Alekseev, 1989), i.e., 2 or 3 m.y. before the impact.

Most frequently, several big asteroids fell successively, thus triggering a series of extinction impulses. Two asteroid impacts took place in the Triassic–Jurassic boundary time (Alekseev, 1998), one in the late Norian (209.6 Ma) and another in the late Rhaetian (205.7 Ma). At least two extinction peaks are recorded in the Permian–Triassic boundary deposits, and five peaks are established in the Cenomanian–Turonian boundary strata (Veimarn *et al.*, 1998). The comet-induced extinctions during the Anisian (241.7–234.3 Ma) and Ladinian (234.3–227.4 Ma) are poorly recognizable against the background of the intense extinction resulted from the asteroid falling.

The specific character of ME events in different time intervals also reflects the fact that the asteroid or comet impact may cause dissimilar consequences in different

Table 3. The Phanerozoic mass extinction (ME) events on the Earth and the moments when the Sun crossed jet streams of the Galaxy

Stratigraphic position of mass extinction event (abbreviations for peaks in Fig. 3a)	ME level	Boundary ages, m.y.	Calculated time, m.y.
Pliocene		5.3–1.8	3
Oligocene–Miocene boundary		23.8	22
Middle Eocene–Priabonian (uEoc)		41.3/37.0	43
Maastrichtian–Danian (Maas)	GME	71.3 ± 0.5/65.0 ± 0.1	66
Cenomanian–Turonian		98.9 ± 0.6/93.5 ± 0.2	90
Barremian–Aptian		127.0 ± 1.6/121.0 ± 1.4	118
Kimmeridgian–Tithonian (Tith)		154.1 ± 3.3/150.7 ± 3.0	150
Pliensbachian–Toarcian (Plie)		195.3 ± 3.9/189.6 ± 4.1	187
Norian–Rhaetian (uNori)	GME	220.7 ± 4.4/209.6 ± 4.1	214
Anisian–Ladinian		241.7 ± 4.7/234.3 ± 4.6	234
Tatarian (Changhsingian)–Griesbachian (Tatr)	GME	252.1/248.2 ± 4.8	253
Sakmarian–Artinskian		282/269	272
Stephanian (Step)		303–290	293
Serpukhovian–Bashkirian		327/323	316
Tournaisian–Visean	(GME)	354/342	340
Frasnian–Famennian (Fras-Fame)	(GME)	370/364	368
Ludlow (Ludl)		423–419	400
Ashgill–Llandovery (Ashg)	GME	449/443	437
Llanvirn–Llandeilo		470/464	464
Tremadoc–Arenig		495/485	484
Dreisbachian		505–495	503
Lenian (Botomian) (uBoto)		524–518	522
Tommotian		534–530	543
The Phanerozoic lower boundary prior to 1993		570 ± 15	566

Note: Slash separates dates for the ages lower boundaries; (GME) great mass extinction.

regions of the globe. The significance of extinction, volcanic activity, climatic changes, and other phenomena in one hemisphere can be weakly pronounced in the other, and vice versa. This is a serious obstacle for correlation of coeval deposits formed during the ME events.

To avoid this obstacle, we should correlate chronostratigraphic boundaries not with the levels of the extinction peaks, but with those marking the taxonomic diversification of animals and plants after cessation of the comet bombarding. This principle was used to establish the Jurassic–Cretaceous, Permian–Carboniferous and many other boundaries. Therefore, there is some lag of several million years between the comet bombardments and relevant boundary markers.

Let us turn back to Table 3. If we take into consideration the aforesaid, the comparison of observation and calculation results shows that the calculated moments, when the Sun crossed the galactic jet streams, are well coincident in time with the empirically determined epochs of mass extinction of living organisms on the Earth.

Hence, the basic supposition of the Keplerian model that the anomalistic period of the Sun motion (250 m.y.) is a multiple of the rotation period of the core disk of the Galaxy (50 m.y.) is correct. However, another supposition that the orbital and anomalistic periods of the Sun motion are equal appears to be incorrect. Otherwise, the ME events of equal intensity would repeat each "galactic year," but the alteration trend of "greater" and "lesser" ME events is different (Table 3).

With due account for distribution of the gravitation potential in the Galaxy, it was figured out that the orbital period of stars of the galactic plane, at the distance of Sun from the Galaxy center, should be 223 m.y., i.e., slightly lesser than the anomalistic period of the Sun (250 m.y.). This circumstance implies the slow turning of the apse line of the solar orbit by $\sim 1.4 \times 10^{-7}$ angular degrees per year toward the direction of the Sun movement.

Figure 3 shows secular variations of marine animals, which became extinct during the Phanerozoic (upper diagram) according to data of Sepkoski (1995), and the Sun position relative to the galactic spiral arms (lower diagram). The arms are tentatively considered as 1 kpc long. It is a general opinion that, simultaneously with the Sun orbital motion, the entire four-armed spiral construction rotates around the center at the angular rate $\omega_\phi = 2.8 \times 10^{-8}$ rad/year that gives the rotation period of 223 m.y. The newly calculated moments, when the Sun crossed the galactic streams, are indicated by numbered points on the solar orbit (Fig. 1).

There is a minor difference between moments of comet bombarding, if they are calculated using models, one taking into account (Fig. 3b) and other discarding (Table 3) the solar orbit rotation. The more precise model shows the lags of 2, 3, and 4 m.y. for orbital points 6, 7, and 8, respectively (Fig. 1). Difference between other points is negligible or absent. Therefore, we can neglect the solar orbit rotation when calculating the moments when the Sun entered the jet streams.

However, that factor is obligatory for determination of time intervals, when the Sun crossed four spiral arms of the Galaxy (Fig. 1). As one can see (Fig. 3b), the "great" ME events took place when the Sun was simultaneously in the jet streams and the galactic spiral arms. In particular, the Maastrichtian–Danian extinction was induced by the Sun position inside Arm III; during the Norian–Rhaetian extinction, the Sun entered Arm I (the Carina–Sagittarius Arm), and the Frasnian–Famennian–Visean extinction is coincident with the Sun movement along the galactic arm IV. Three significant extinction events at the Tatarian–Griesbachian and Ashgillian–Llandoverian boundaries and during the Early Ordovician mark moments of Sun residence in Arm II (the Perseus Arm).

Our calculations show that we should add the Jurassic–Cretaceous and Carboniferous–Permian boundary events to six "great" ME of the Phanerozoic times, though they do not correspond to the latter by the extinction intensity. The relatively low extinction peaks at two former boundaries (Fig. 3a) can imply that no big asteroids, which can cause mass death of organisms, fell on the Earth in these time intervals.

Another reason, which seems similarly important, is the difficulty in correlation of boundary deposits. The ambiguous correlation is known to result in erroneous evaluation of extinction levels. For example, Alekseev (1998) concluded that evaluation of the Jurassic–Cretaceous extinction is dependent on the uncertain position of the Tithonian and Volgian stages, which used to be considered coeval. However, new data substantiating the Berriasian age of the upper Volgian deposits have been recently obtained (Sei and Kalacheva, 1997), and the National Interdepartmental Stratigraphic Committee lowered the Jurassic–Cretaceous boundary down to the base of the upper Volgian Substage.

CONCLUSIONS

Leaving aside all problems concerning substantiation and reliability of available geochronological scales, we would like to state that there is certain correlation between the geologic time scale, mass extinction events, and cosmic phenomena in the Galaxy and Solar System.

The considered data allow us to suggest the following:

(1) Since origination of the Earth and other planets of the Solar System, they have been impacted by big cosmic bodies, such as asteroids and galactic comets. Impacts of these bodies caused serious natural catastrophes on the Earth, records of which could not escape revealing by recent geological investigations of the Phanerozoic deposits to be interpreted as stratigraphic boundaries of variable rank. The moments of most intense comet bombarding correspond to boundaries between periods (systems), and those of lesser intensity characterize boundaries between epochs (series). Impacts of individual asteroids are detectable in geological records as boundaries between ages (stages).

(2) The galactic comets fall on the Earth only in the periods, when the Sun crossed the jet streams of the Galaxy. These periods were as long as several millions of years and repeated each 20–37 m.y. Every time when the Sun was inside the streams, approximately 10^4–10^5 galactic comets of nearly 10^{13}–10^{18} g in mass may fall on the Earth. The middle-sized comets could be of one or two kilometers in diameter, having the mass of ~10^{15}–10^{16} g. In average, one galactic comet reached the Earth approximately each 100 years, when the Sun was inside the stream. In dependence on the Sun position on the galactic orbit, the impact frequency could vary by one to two orders of magnitude.

(3) In contrast to asteroids, galactic comets collapsed completely in the atmosphere and affected the Earth surface as powerful impact waves. Creating no craters, the impact waves caused colossal destruction and mass extinction of living organisms in a vast territory. A middle-sized galactic comet is able to release in the Earth atmosphere as much energy as ~10^{24} J that is hundreds of million times greater than the explosive energy of the Tunguska cosmic body.

(4) Nearly 200 Phanerozoic events can be interpreted as the impact results of solitary asteroids more than 3.5 km in diameter, which accidentally reached the Earth with periods of 2.9 m.y. in average. About 45% of their impacts left signature at the levels of age (stage) boundaries. Other impacts (55%) probably took place in the periods of comet bombarding, but some could be recorded as boundaries of finer geochronological units (chronozones).

(5) The falling asteroids with critical diameter of ~3.5 km have the energy of ~10^{22}–10^{23} J. The impacts of greater energy could have global consequences that lasted for a long time. More frequent falling of lesser (lower-energy) asteroids affected a restricted number of

regions. They were of regional or local significance, and corresponding events could mark boundaries of minor stratigraphic subdivisions. Many of these events left no records. This means that we cannot elaborate a standard time scale for the Phanerozoic, the resolution of which would exceed the stage range.

We understand that the problems touched in this work need to be analyzed and discussed much more comprehensively. However, it is clear now that the deciphering of many geological events in the Earth history will be incomplete and even erroneous without taking into account the impacts of cosmic phenomena in the Galaxy on the Solar System.

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Problems of Perfecting Zonal Stratigraphic Scales

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Abstract—Problems of perfecting zonal stratigraphic scales based on the paleontologic evidence are discussed. In all the diversity of biostratigraphic units we operate with, there are two basic types—biostratigraphic zones lacking strict definition of their boundaries and biochronological zones with strictly defined boundaries. Correspondingly, peculiarities of biostratigraphic and biochronological scales and methods of their modernization are considered. When a zone is used as a unit of a section and of a reference scale, the negative consequences are unavoidable. The higher correlation potential of a given scale can be achieved by widening the paleontological characteristic of its zonal units. The chronological continuity of zonal units is determined by the evolutionary succession of fossils used as a basis for the scale elaboration. The correlation accuracy increases when the scale resolution grows. Introduction of non-independent chronological markers into biochronological scales may preserve, but never increases and usually reduces the correlation accuracy. The higher resolution of the biochronological scale can be achieved, provided that its universality and correlation potential are maintained, using the same means that were applied for its elaboration.

Key words: biochronological scale, biostratigraphic scale, zone types, stratigraphic zone, correlation accuracy, scale detailing.

INTRODUCTION

Application in stratigraphy of zonal chronological scales substantiated paleontologically is aimed at solution of two problems: dating and correlation of distant stratigraphic objects. Perfection of scales implies the increase in their resolution and correlation potential. When posing and solving these problems, one should take into consideration the dual character of the scale properties: the better detailed scale has a lower correlation potential.

Of a particular importance is seeking for the continuous chronological scales that is one of the aspects of their perfection. To set on a constructive discussion of perfection methods, it is necessary first to decide which type of the scales we want to be perfected. In addition, one should unambiguously discriminate between zones usable as the scale units or as subdivisions of a section prior to detailing them. Finely, it is also important to understand difference between the stratigraphic and chronological successions of events, which are used in stratigraphy for elaboration of chronological scales. All these topics are discussed below.

TYPES OF ZONAL SCALES

The Stratigraphic Code (*Stratigraficheskii kodeks...*, 1992) recommends to use two basic types of zonal units of the so-called complex substantiation—chronozones and local zones, in other words, two types of scales, which differ from each other only in the correlation potential. The former are thought to be of a global significance, the latter are provincial. In my opinion (Chernykh, 1995), every scale constructed correctly can potentially be universal and, after imparting the necessary paleontological characteristic to zonal units, limits of its use can be widened as one needs.

Principal differences between two scales are determined by their basis. From this point of view, there are two types of chronological scales, which are based on paleontological principles: the scales based on the evolutionary succession of fossils and the scales based on stratigraphic distribution of certain fossil assemblages (associations) in the particular sections. The last type that can be termed biostratigraphic precisely corresponds to local and (chrono)zonal scales, as they are defined in the Stratigraphic Code. I used the term "biochronological" for the scales that are based on the evolutionary succession of fossils, though it can be considered etymologically lame, creating an impression that biochronological scales (BCHS) are opposed to biostratigraphic ones (BSS) as independent of stratigraphic criterion. In fact, it is not the case, and the necessity of the strict stratigraphic control in elaboration of any stratigraphic scale is repeatedly emphasized in further discussion.

Let as characterize biostratigraphic and biochronological scales for clarifying difference between them. All the detailed units defined as an assembly of beds within a certain stratotype section and having paleontological characteristic can be considered as "biostratigraphic zones." The chronological succession of such units can be considered, in turn, as the biostratigraphic

scale. Zonal units of the biostratigraphic scale represent a stratigraphic succession of fossil assemblages, chronological boundaries between which are ambiguous. Moreover, the postulated evolutionary succession of fossil assemblages increases this uncertainty. Actually, the corresponding boundaries are often inside transitional intervals, and different researchers can place them at different levels (*International stratigraphic...*, 1976).

Units of the biochronological scales are distinguished directly in the scale and strictly correspond to elementary evolutionary events (appearance/extinction of taxons belonging to a single lineage). Using the biochronological scales with unambiguously determined boundaries of zonal units, one cannot voluntary place the boundaries in sections. In this case, the methodical approach to applying the zonal biochronological scales in correlation becomes particularly clear. Usually, boundaries of the correlated stratigraphic unit do not correspond to zone boundaries in the scales. They can be correlated tentatively, based on their spatial proximity in the section with the only purpose—to correlate a stratigraphic unit throughout the region of the maximum possible dimensions. Of course, It would be groundless to include such zones into the hierarchy of stratigraphic units. Moreover, it is clear that their recognition in the section is not an ultimate goal, but an intermediate preparatory operation that is necessary only for correlation of boundaries of stratigraphic units (stages, horizons, members).

Further, I comment seven types of zones mentioned as biostratigraphic in the Stratigraphic Code (*Stratigraficheskii kodeks...*, 1992, p. 45; see Fig. 1 quoted from the code).

The taxon-range zone is a group of strata corresponding to the total stratigraphic range of a fossil taxon (Fig. 1, *1*).

In every particular section, stratigraphic distribution of a taxon represents its teilzone, and one should determine the teilzones in every section and then to summarize them in order to define the complete stratigraphic range of the taxon. This can be done only with the help of a chronological scale, where the taxon distribution boundaries in all examined sections are marked. Inasmuch as only the biochronological scale is appropriate for establishment of exact position of the teilzone boundaries, the biozone is a derivate of zonal units of the biochronological scale, based on which the teilzone position was determined. Thus, biozones cannot be defined in the section until their position is determined in the biochronological scale. Consequently, the biozones are of the auxiliary significance in definition of complex characteristic of zonal units in the biochronological scales and cannot be considered as basic (independent) zonal units.

The concurrent-range zone spans the coincident parts of stratigraphic ranges of two selected taxons (Fig. 1, *2*).

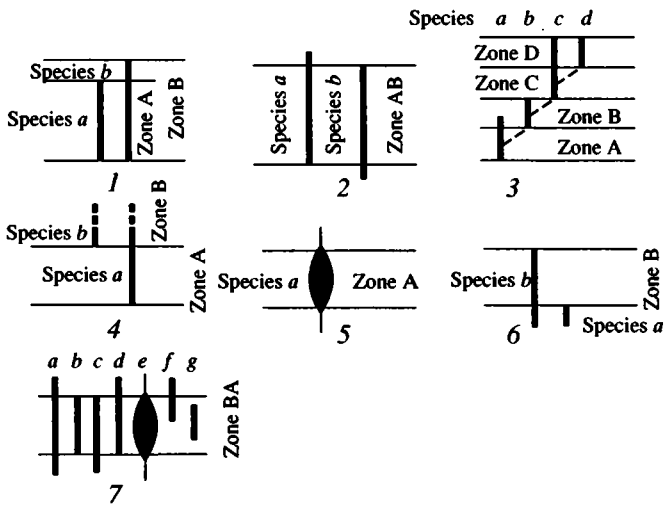

Fig. 1. Biostratigraphic zones (*Stratigraficheskii kodeks...*, 1992):
(*1*) taxon-range zone (biozone); (*2*) concurrent-range zone; (*3*) phylozone; (*4*, *6*) interval zone; (*5*) acme zone (epibole); (*7*) assemblage zone.

When we are dealing with the overlapping interval of biozones of two taxons, its boundaries are determined similarly to those of biozones, and the interval of concurrent occurrence is defined as that of one or several units in the scale used for recognition of biozone boundaries. In this case, the concurrent-range zone is of a auxiliary significance too, and represents, similarly to the biozone, a derivate of the biochronological scale units.

When a scheme shows the concurrent occurrence range of two species in the particular section, one should remember that this range is always determined within limits the concrete boundaries that are preliminary determined by the independent method, par example, by lithology (occurrence of two taxons in a single bed, member, formation). In this case, we deal with a classical example of the biostratigraphic zone, i.e., with an assembly of beds having concrete paleontological characteristics (concurrent occurrence of taxons).

Hence, of the concurrent-range zones cannot be referred to units of biostratigraphic or biochronological scales until the definition presented above is specified.

The lineage zone corresponds to the stratigraphic range of a taxon (taxons) representing members of a given evolutionary lineage or phylogenetic trend (Fig. 1, *3*).

Figure 1 demonstrates principles of recognizing the lineage zones using the criterion of "appearance" and "extinction" of individual species belonging to a single lineage, but the sketch likely represents a particular section, because "beds" are indicated. Consequently, the lineage zones of the section represent a result of using the previously constructed BCHS. As I mentioned earlier (Chernykh, 1995), the essence of biochronological

scale units is biological, and it would be more correct to define the lineage zones as units of the biochronological scale, where their boundaries coincide with the appearance levels of successive index species. The phylogenetic basis is most reliable for establishment of strict boundary positions in the biochronological scale and continuity of the latter. The lineage zone that is defined using appearance levels of successive species of a single phratria represents a zonal unit of the biochronological scale.

The interval zone includes strata of a continuous succession between the first appearance level of some characteristic taxon (usually index species) of the considered zone and first appearance level of the taxon (usually index species) characteristic of the overlying zone (Fig. 1, *4*). The interval zones can also be defined as beds bracketed between the extinction levels of taxons characteristic of the given and underlying zones (Fig. 1, *4*, *6*).

When successive index species belong to a single evolutionary line, the interval zone is identical to the lineage zone and can hardly be considered as having an autonomous significance. When index species belong to different lineages, the interval zones can be correctly defined only using the known relationships between biozones of these species, and we collide again with the problem of biozone definition.

In the scale calibration, the criterion of "appearance" has some advantages as compared with that of the "extinction." The "species appearance" event is detectable in the paleontological record based on the lowermost find of the fossil taxon in a section, whereas the "species extinction" event can be recognized using data on many sections. In addition, the "extinction" criterion is ineffective in the case of redeposited fossils (Chernykh, 1995).

The acme zone (epibole) represents strata where a particular taxon is most abundant (Fig. 1, *5*).

In other words, it corresponds to the stage of maximum development (flourishing) of a taxon. This stage is, however, facies-dependent and qualitatively uncertain, connected via gradual transitions with previous and subsequent stages of the taxon development. Hence, the adequate (in terms of strictly defined boundaries) positioning of the acme zone in the biochronological scale is impossible. The acme zone represents an indicative characteristic of some units of the biostratigraphic scale, though with corresponding limitations for its wide application.

The assemblage zone is an assembly of beds characterized by a fossil assemblage different from those in underlying and overlying beds (Fig. 1, *7*).

As evident from definition, this is a typical biostratigraphic zone. Figure 1 clearly demonstrates that boundaries of the assemblage zone can be placed at different levels. For instance, in addition to the version shown in the guide, other variants of zone boundaries are possible: the lower boundary can be lowered to the appearance level of species *c*, *f*, or *g*; the upper one, to the extinction levels of species *d*, *f*, or *g*. These variants do not violate the definition mentioned above and illustrate the uncertainty of boundaries of the assemblage zones.

On the other hand, considering the figure, one can recognize the *concurrent-range zone* of species *c* and *d* that represents a particular case of the assemblage zone. Thus, only two types of zones can be objectively defined (i.e., based on principles of elaboration of a zonal scale). The zone of the first type is biostratigraphic; its essence is defined as an assembly of beds subsequently characterized in paleontological aspects. The biochronological zone (second type) is determinable within limits of elementary evolutionary events, such as "appearance/extinction" of species, thus being biological (paleontological) in essence. Methods of perfecting the biostratigraphic and biochronological scales are discussed below.

ZONE AND STRATOZONE

Until recently, there is no terminological discrimination between zonal units sections and scale: in both cases, the same term "zone" is in use.

The indifferent usage of the term "zone" with respect to a scale (further simply "zone") or a section unit (further "stratozone") is not incidental. It is rooted in the widely accepted concept that precisely the stratified rock sequence (either an individual, or composite section that is assembled from individual ones) represents the stratigraphic scale (Ganelin, 1992; Naidin, 1994). To illustrate such scornful relation to the problem of discrimination between the zone and stratozone, the last domestic Stratigraphic Code (*Stratigraficheskii kodeks...*, 1992) should be mentioned. All definitions of different-type biostratigraphic scales begin here with words "strata," "group of strata" or "assembly of rocks" (p. 45) and nothing is said about zonal units of biostratigraphic scales.

As shown below, that terminological deficiency in differentiation of the "zonal unit" notion has negative consequences for posing and consideration of many important points in zonal stratigraphy.

(a) Stratigraphic and Chronological Succession of Events

According to a just remark by Meyen (1974, p. 14), "in their essential theoretical principles, all stratigraphic scales are identical, chronological in nature." Nevertheless, this does not mean that the stratigraphic succession of events can serve as a chronological basis. There are essential differences between stratigraphic and chronological successions of events used to substantiate the stratigraphic scales.

Constructing any chronological scale that is paleontologically substantiated one begins with stratigraphic subdivision of a section. The stratigraphic succession of

beds in the section serves as a time axis for events and processes distinguishable in the geological record. Zonal scales based on paleontological data inherit the general chronological trend from the stratigraphic succession of beds, although the principle of calibrating scales is actually independent of principles used for section subdivision into lithological units. Defining any assemblages or groups of fossil taxons confined to certain parts of the particular section, or registering the appearance and extinction levels of different forms, we can "attach" them to certain intervals or make coincident with some lithological boundaries. After these operations, we have an impression of the constructed scale corresponding to the studied section that is subdivided into units based on paleontological data. However, this impression is far from being correct.

Several additional operations should be performed to convert the stratigraphic succession of paleontological events into a real geochronological scale. First, we should solve the problem of selecting criteria of positioning the unit boundaries in the scale, because not all the features used for recognition of the boundaries in the section are suitable for this purpose. Further, we should be sure that the stratigraphic succession of events that we select to mark the unit boundaries in the section is chronologically continuous, but we cannot answer this question observing only spatial relationships between zonal units defined in the section. In addition, we should be sure that paleontological events, established in the particular section and accepted as a basis for scale construction, are of sufficient correlation potential. To meet all these requirements, we should accomplish a complex study to extract the complete paleontological information in its, so speaking, pure form for the subsequent thorough analysis. The biochronological scale, which should crown these efforts, will be so different from the primary draft version that an illusory idea of the sequence-based scale will completely disappear. Only after the final construction of the scale, we can pass from stratigraphic subdivision of the section to its chronological calibration using this scale and, then, to correlation.

(b) Boundaries of Zonal Units

Let us consider an old problem of criteria defining the zonal boundaries, in particular, the criteria of "appearance/extinction" levels of species taxa. It should be noted first that these criteria are termed from the standpoint of the "substrate" basis of zonal units, because they concern the first appearance and extinction of fossil forms in the section. Exactly from this standpoint, disadvantages of criteria, which are markers of zonal boundaries, have been considered: it was noted that "the first appearance" can be in fact not the first one and "extinction" not the last one.

When constructing a zonal scale, we actually deal with the stratified section, inferring the required chronological succession of taxa from their "first appearance" and "extinction" in that section. The probable incorrectness in defining the "first appearance" and "extinction" levels of fossils influences insignificantly the correctness of conclusion on the chronological species succession that is repeatedly controlled in different sections. It should be emphasized that we do not define zones in the section, when it is under study prior to construction of a zonal scale; we obtain only the paleontological substantiation of their position in the scale. When the stratigraphic succession of the appearance and extinction levels of fossils in the section is established, we interpret it as a chronological succession of evolutionary "appearance" and "extinction" events of corresponding taxa. Precisely the qualitative certainty of elementary evolutionary events responsible for "appearance" and "extinction" of species means that they can be used for the strict (unambiguous) positioning of zonal boundaries in the scale.

When the scale is ready, we again turn to the section and, using this scale, subdivide it into zones assuming that the lowermost finds of successive index species (fossil forms) approximately coincide in the section with the "appearance" events of these species and mark boundaries of zonal units. Accompanying inaccuracies in determination of "first appearance" levels have nothing to do with our correctness by the scale construction and concern only the applied operation. In this connection, it should be noted that any scale leads in practice to ambiguous results and is always supplemented with indications of the measurement accuracy, calculated uncertainty limits, etc. However, the fact that we discuss not the faulty criteria used by the scale creation, but the way to overcome incorrectness of the scale application, diverts the discussion. For instance, it is likely that a sufficiently complete paleontological characterization of zonal units reduces the extent, to which position of zonal boundaries in the section depends on fluctuations of "first appearance" levels of zonal index species.

When the unambiguous determination of zonal boundaries in the scale is not obligatory, we can be satisfied only with the chronological succession of events easily recognizable in the section. This condition is met by stratigraphic succession of biostratigraphic zones (Oppel-zones, assemblage zones, concurrent-range zones), every of which is characterized by "an association or aggregation of selected taxa of restricted and largely concurrent range, chosen as indicative of approximate contemporaneity" (*International stratigraphic...*, 1976, p. 58). For instance, boundaries of the Oppel-zone are determined, as is accepted, by the stratigraphic range of "the ensemble of fossil forms considered distinctive of the zone" or as "the outer limit of the concurrence of the selected taxa designated as diagnostic" for the concurrent range zones. It is clear that these boundaries can be shown in the scale only when "distribution limits" of diagnostic taxa are determined in some independent scale. Most frequently, the latter is a stratigraphic succession of beds in the partic-

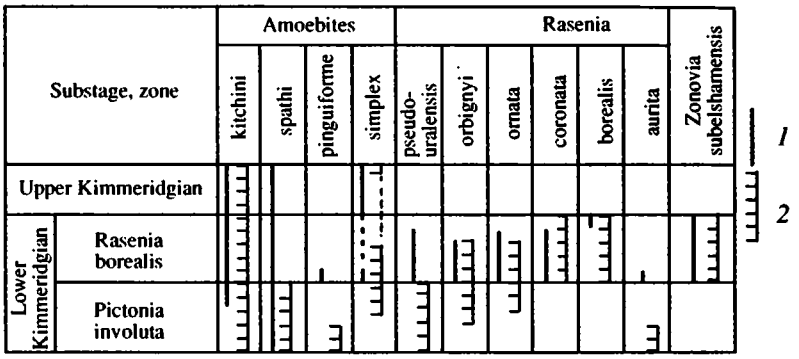

Fig. 2. Distribution of some characteristic lower Kimmeridgian ammonites in the Taimyr Peninsula (*1*) and North Urals (2) (after M.S. Mesezhnikov in Stepanov and Mesezhnikov, 1979; Fig. 5.58).

ular section, i.e., the simple lithological scale. This creates an illusion (when paleontologic characteristics of defined lithological units are known) that we have a biostratigraphic scale with distinct boundaries between zones. In fact, distinct are boundaries in the applied lithological scale, which cannot be inherited by the biostratigraphic scale based on the succession of fossil assemblages. This becomes clear, when we turn to some other sections where position of zonal boundaries are established, as was mentioned, voluntarily within a certain interval.

Simultaneously, the correctly established succession of such assemblage zones can serve as a scale for dating and correlation of a similar succession of stratigraphic units. It cannot be used for correlation of relevant boundaries, because their determination is beyond the scale capability, and they can be placed only approximately "between assemblages."

(c) Zone as Elementary Unit of the Scale

To continue comparison of zonal units and corresponding subdivisions of a section, let us dwell on another essential point. In every particular section, we deal with teilzones, which are, similarly to the section itself, also unique, and correlation based on biochronological scale is always potentially inaccurate owing to this circumstance. To visualize the scale inaccuracy, we should introduce a condition of a minimal range of zonal units in the scale (Chernykh, 1995). In any case, the zonal scale does not inform us about completeness of zones in the section. In practice, we always correlate sections at the "zone level" of accuracy regardless the actual position occupied by the particular object in the section. In this connection, some frequently occurring terms, such as "lowermost" and "uppermost" parts of a zone, represent unjustified attempts to consider the scale as identical to section. For instance, it is unclear how it is possible to establish the confinement of some early Kimmeridgian ammonites to certain parts of zonal units of the ammonite-based scale (Fig. 2), if one compares their stratigraphic ranges in the North Urals and Taimyr (Stepanov and Mesezhnikov, 1979). Moreover, the occurrence level of zonal species *Rasenia borealis* in the Taimyr Peninsula is shown in the uppermost part of the synonymous zone (Fig. 2). I am quite sure that such facts can be established only by using a more detailed scale, but cited authors mention nothing of this kind. Most probably, this is a frequent case of scale and section identification.

Zones of the chronological scale represent chronologically uniform, equivalent minimal units. When somebody speaks about the "basal," "middle," or "uppermost" parts of a zone, there should be some material criteria of their recognition. If these criteria meet requirements used to establish zonal boundaries in the scale, the latter should be divided into smaller units (autonomous zones). These new zones should be considered as minimal units of the scale, virtually instantaneous and indivisible from the geological standpoint.

A possibility to distinguish parts of a zone appears when we turn to the stratozone, i.e., to the particular stratified section, where position of zonal units is established using the biochronological scale. In the substrate equivalent of the zone, i.e., in the stratozone, we can define its lower, middle, and upper parts using lithological criteria. However, as was mentioned, the biochronological scale lacks information indicating which part of the considered zone is represented in the section. When we decide to subdivide stratozones based on criteria that differ from those used to distinguish zones in the scale, we should realize that position of new markers in the scale will be uncertain, and that defining smaller parts of a zone, we begin, in fact, to create a new scale on the new base. This problem will be considered further when discussing ways of detailing the biochronological scales.

PERFECTION OF THE SCALE

(a) The Correlation Potential Increase

Universalization of biostratigraphic and biochronological scales develops in a similar way—by progres-

sive enhancement of complex characteristics of their zonal units. Nevertheless, the results are highly dependent on the type of scales we universalize—biochronological or biostratigraphic.

For instance, correlating the increasingly remote sections, we enlarge uncertainty of zonal boundaries determined with the help of biochronological scale, and this leads to the uncontrolled quantitative increase of taxa considered as diagnostic for a zone. In addition, synonymous zones are frequently defined in different sections using some species only, but not the entire diagnostic assemblage. This blurs the zone position in the section, and inclusion of endemic forms into the diagnostic assemblage further aggravates consequences of broadened complex characteristics of zonal units. The iterative character of stratigraphic procedure (when the results of correlation are used for correction of the scale) leads in this case to appearance of pseudouniversal scales, because the number of fossils in the zonal assemblage becomes tremendous and facilitates recognition of the corresponding zone in any section, but at different levels in accord with conclusion of different researchers. This is particularly true of zones with microfauna that is characterized by a great number of species and subspecies, many of which do not occur in each section, and the artificial aggregation of these fossils opens a wide opportunity for voluntary selection of the diagnostic zonal assemblage. Something like this happened with the fusulinid zonal scale that was initially created for the Carboniferous–Permian boundary interval in the Russian platform and then used for distant correlation with Uralian and Tethyan sections. Inasmuch as even index species play frequently a nomenclature role in this scale and can occur in several contiguous zones, the objective test of zonations, which are suggested by different specialists for the same deposits, appears to be impossible. Substitution of the assemblage base with the phylogenetic one, i.e., transition from the biostratigraphic scale to the biochronological one that was initiated recently (Davydov, 1988), is probably the only way to solve the above problem.

The Carboniferous–Permian conodont-based zonal scale is also under way to such a transformation. At the first stage of its elaboration, the established conodont assemblages were correlated with zonal units of the fusulinid biostratigraphic scale (Chuvashov *et al.*, 1983), and the chronological succession of these assemblages was traced. At this stage, conodonts only supplemented the paleontological characteristic of fusulinid zones, neither changing their resolution nor eliminating the uncertainty in position of zonal boundaries that is peculiar of biostratigraphic scales.

The further study of the conodont morphology and stratigraphic distribution revealed a distinct morphological trend in development of late Gzhelian and Asselian streptognathids: *Streptognathodus wabaunsensis–S. isolatus–S. glenisteri–S. cristellaris–S. sigmoidalis–S. constrictus–S. postfusus*, and enabled elaboration of the zonal biochronological scale based on the appearance levels of these species (Chernykh and Reshetkova, 1988). This scale exceeds the resolution capability of the initial fusulinid scale and, unlike the latter, is characterized by strict positions of zonal boundaries. The basic succession of most above-mentioned species is now established, in addition to the Urals, also in North America, and the conodont scale enabled global correlation of the Carboniferous–Permian boundary (Chernykh and Ritter, 1994). Thus, there are two zonal scales of the Carboniferous–Permian boundary interval: fusulinid-based biostratigraphic and conodont-based biochronological.

Universalization of biochronological scales with distinct zonal boundaries (appearance levels of successive index species) presumes enhancement of zonal paleontological characteristics and is achievable by determination of chronological distribution of corresponding species (and any other markers) with the "zone"-range accuracy, provided that zone is the elementary chronological unit of the scale. Species whose distribution in all examined sections is limited by zone boundaries are introduced into the zonal characteristic as index species of the corresponding unit. The zonal assemblage progressively growing in volume can determine a zone even if its index species are absent. It should be noted that broadened paleontological characteristic of zonal units does not increase the resolution degree of biochronological scales.

(b) Assurance of the Chronological Continuity of the Scale

It should be emphasized that not all researchers consider the chronological continuity of successive zonal units as an obligatory property of the correct biochronological scales. Their main argument is that biostratigraphic data themselves are discontinuous. Sometimes, this is considered as a temporal obstacle for recognition of zones in poorly studied regions (Bodylevskii, 1964), or as a reason to discard the idea of continuous scales (Guex, 1991). The latter work is dedicated solely to discrete zonal scales, the units of which are characterized by unique and mutually exclusive species assemblages. Such units are non-contiguous, isolated from each other by separating intervals. Guex believes also that the traceable succession of the appearance (or extinction) levels of fossils are of subsidiary significance for defining zonal units of this kind, which are termed the "unitary associations" and represent the assemblage zones in their essence. This is probably correct until the moment when the scale should be applied to another section. However, as soon as there is a need to examine the other sections, it becomes clear that we cannot ignore the "subsidiary" events of "appearance" or "extinction" when defining boundaries of discrete zones. This is well seen in figures presented by Guex in his work, for instance, in his Fig. 1.1, reproduced here (Fig. 3). In this scheme, boundaries between discrete

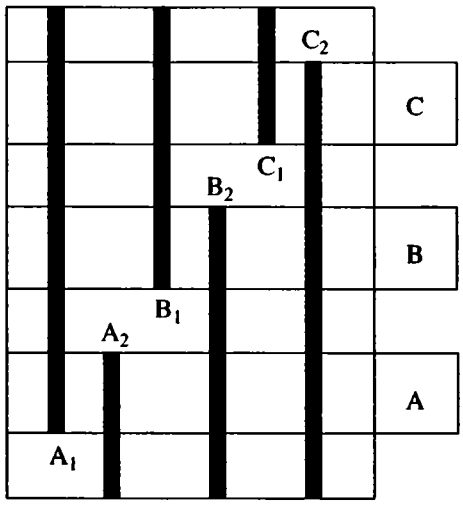

Fig. 3. Discrete zonation A, B, C based on data on hypothetical stratigraphic distribution of species A_1, A_2, B_1, B_2, C_1, and C_2 (Guex, 1991; Fig. 1.1).

zones (A, B, C) correspond precisely to the appearance and extinction levels of corresponding species. Using this criterion and the same factual materials, one can suggest a continuous scale consisting of seven zonal units. The discrete biochronological scales (i.e., scales with strictly positioned zonal boundaries) can be obtained, in fact, when we rarefy units of continuous scale that may be a useful step for increasing of their correlation potential. When the exact position of boundaries is unknown, we receive a discrete biostratigraphic scale, units of which are defined only by a joint occurrence of certain species (Oppel-zones, assemblage zone, and others). Thus, solution of the continuity/discontinuity problem can hardly be avoided in the course of biochronological scale elaboration.

An assurance of the biostratigraphic scale continuity is usually understood in terms of nearby zones adjoining (Rauser-Chernousova, 1967). It should be understood, however, that the continuous occurrence of fossils and possibility to distinguish a continuous succession of assemblage zones in a particular section do not ensure adjoining of zones in the chronological sense. For instance, when the boundary between contiguous zones corresponds to a hiatus or is facies-dependent, the same zones can be separated in other sections.

With respect to the biochronological scale, the principle of zones adjoining should be replaced by the principle of the scale chronological continuity. I think that this is a most serious requirement, and the criterion of successive zonal assemblages that was frequently mentioned in this connection is insufficiently effective to satisfy it. Most suitable for calibration of continuous zonal biochronological scale is the evolutionary (phylogenetic) basis. Distribution of successive species of a single phratria that obeys the law of divergent evolution is characterized by intervals of overlapping (Fig. 4) and secures the continuity of the scale constructed on this basis. The fact of overlapping distribution of contiguous index species in the section ensures the scale continuity even if some members of the phratria are missing.

There are different views on importance of lineages in stratigraphy, but all researchers agree that zonal units, if they are distinguished based on evolutionary changes, are warranted from repetitions and are facies-independent. Irreversibility of evolution is indeed useful for selection of the chronologically successive index species. In principle, the necessary chronological succession of several species of one or several genera can be selected empirically, based on their stratigraphic relationships and regardless the phylogeny. Such an approach to zonal biostratigraphic scales was suggested by Rubel' (1982) but neither he, nor anybody else explained so far how can it be done without phylogenetic base, and what can be used instead for recognition of continuous succession of zonal units in the biochronological scale. I believe that skeptical evaluation of phylogenetic lineages because of their subjectivity is unjustified. Recognition of lineages should strictly be controlled by stratigraphic data. Only in this case, the distinguished chronological successions of taxa are at least as subjective as a succession of zonal species based on their stratigraphic position in the section. Simultaneously, the phylogenetic base is of the organizing significance for sampling and investigation of fossils because it calls attention to a purposeful search of forms of one or several concurrent evolutionary lineages. It is impossible to avoid certain theorization using lineages for subdivision of biochronological scales, and therefore the stratigraphic control over selection of chronological succession of index species should be especially strict. In particular, the elementary (three-zoned) biochronological scale must be inferred from a single or several closely spaced sections (Chernykh, 1995).

Thus, the continuity of biostratigraphic scales can be achieved empirically, by defining the additional assemblages, in sections where the known zones are separated by hiatus. The control criterion in this case is the assemblages succession.

The continuity of biochronological scales is assured by the purposeful searches of unknown members (species, subspecies, morphotypes) in the evolutionary succession of taxa used as a basis for the scale elaboration. If the evolutionary succession of forms shows a certain trend of particular feature development, we can establish much easier positions of new forms added to the succession. Thus, it can be stated that morphological trends represent the most suitable basis ensuring the continuity of biochronological scales.

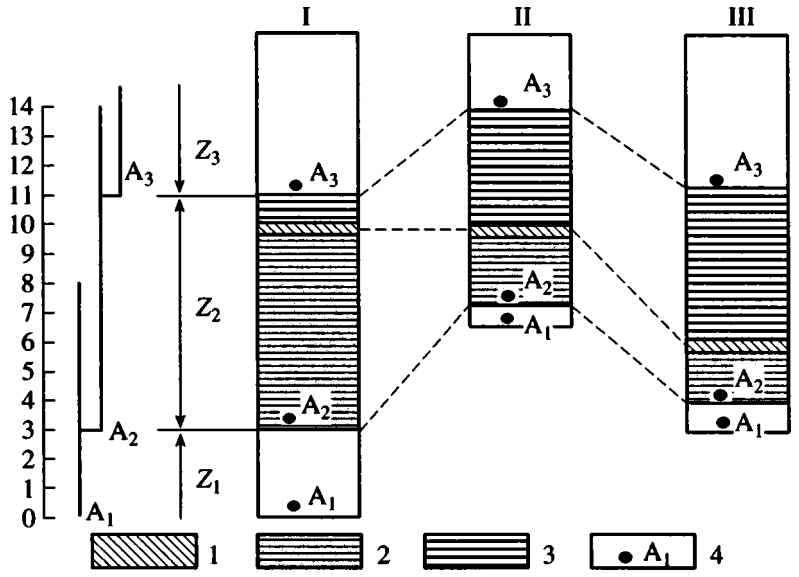

Fig. 4. Correlation accuracy decrease in the case of use of the interzone lithological marker: (1) bentonite bed; (2) part of the zone Z_2 below tuffaceous bed; (3) part of the zone Z_2 above tuffaceous bed; (4) first occurrence levels of species belonging to the genus A. (A_1, A_2, A_3) evolutionary succession of species belonging to the genus A; (Z_1, Z_2, Z_3) zonal units of the biochronological scale; (I–III) sections under correlation; shown in the left is the conditional chronological scale.

(c) Enhancement of Correlation Accuracy of Biochronological Scales

Before discussion of this topic, it is necessary to explain what should be understood under the "accuracy" of correlation based on the biochronological scale. We must remember that correlation of objects with the help of biochronological scale means their chronological identification, i.e., the object reference to one or several zonal units of the scale. Because of the elementary status of zonal units, any object can correspond only to the integral group of zones. In this case, we pay no attention to the fact that the object boundaries find different positions within zonal units of different sections, and two correlated objects are considered as chronologically identical, when they fall in the interval spanning the same number of zones. It is clear therefore that the correlation accuracy is determined by the resolution degree of the biochronological scale and cannot be more precise than the zone range.

The issue of the admissible error in biostratigraphic correlation is mentioned in one of the works by Khalfin (1972). Despite the fact that his conclusions accord with the aforesaid, his approach to solution of the problem deserves a brief comment at least. As a correlation tool, Khalfin considers not the scale but a distribution zone of some archistratigraphic species N (biozone). Only phyletic evolution of species successively transforming into each other (the hypothetical process never observed in the real paleontological material) can probably serve as the basis for the corresponding scale consisting of several biozones. Such a chronological scale would resemble the royal dynasty ("The King dead. Vivat rex!"), but the actual paleontological records point to the overlapping of successive species ranges in the phyletic line. Subdivision of the biochronological scale by appearance levels of successive index species with partly overlapping ranges (Figs. 4, 5) ensures the scale continuity, on the one hand, but reduces its correlation accuracy, on the other. For instance, when the index species A_2 and successive index species A_3 appear in one section (I) earlier than in another one (II), the correlation of a particular stratigraphic unit (horizontal hatching in Fig. 4) can be performed with an accuracy corresponding to the integral "zone + overlying zone" interval. This is, in fact, a maximum error that is admissible in correlation with a help of the biochronological scale. At the same time, the above statement that the maximum accuracy of biochronological scale corresponds to its minimal unit (zone) remains valid.

It is necessary to clarify one point in the approach used by Khalfin to substantiate the fact of the heterochronous appearance of the archistratigraphic species (index species in my interpretation). I mean the external dimensionless "time axis" in the scheme by Khalfin and a conditional time ruler in my own (Figs. 4, 5), though the difference does not change essence of the approach. Inasmuch as natural chronometric scales are unavailable, both theoretical schemes serve only as a mean to illustrate possible errors ("limits of admissible inaccuracy," according to Khalfin) in correlation with the help of biochronological scales. Nevertheless, one should not get an illusion that we can control and assess the chronometric accuracy in correlation by the biochronological means.

The aforesaid suggests that detailing of scales increases the accuracy in correlation on the biochrono-

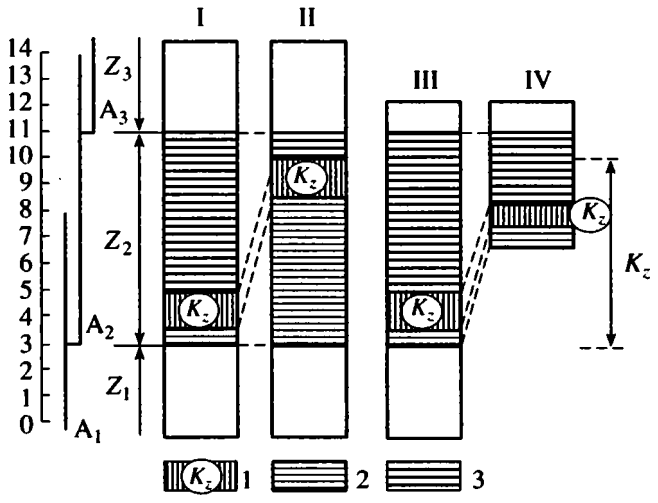

Fig. 5. Correlation accuracy decrease resulting from section subdivision based on polytaxon assemblages: (1) assemblage zone K_z; (2) lower part of the zone Z_2; (3) upper part of the zone Z_2.

(A_1, A_2, A_3) evolutionary succession of species belonging to the genus A; (Z_1, Z_2, Z_3) zonal units of the biochronological scale; (I–IV) sections under correlation; shown in the left is the conditional chronological scale; shown in the right is a total range of the assemblage zone K_z.

logical base. The more detailed is the homotaxial succession of chronometric markers in the biochronological scale, the greater is opportunity for chronological identification of progressively smaller units in the section. Only the rank of units in the correlated sections can serve as an objective indicator of the chronological accuracy of the scale. Nevertheless, even the most detailed ("precise") biochronological scale based on the orthostratigraphic group does not solve, contrary to the widespread opinion, the problem of "establishing and tracing the isochronous levels" (Stepanov and Mesezhnikov, 1979, p. 363). This is a prerogative of chronometric scales, which are still unavailable in stratigraphy, and, consequently, it is impossible to interpret in terms of time the appearance and extinction levels of correlative fossils in different sections.

The available zonal scales are chronological and they can help only to make a chronological identification of different events reflected in the geological record. Nevertheless, even such identification can be performed with the accuracy corresponding to the elementary unit of the scale, in our case, to the zone.

(d) Enhancement of the Biochronological Scale Resolution

Now, I would like to demonstrate that the correlation potential of the biochronological scale could be increased, without changing its universality and correlation accuracy, by the same means that were initially used to create the scale.

As was noted, the best markers of zonal biochronological scales are levels of the successive appearance of species belonging to the same lineage. Precisely in this case, the markers differ from each other in some peculiarities that determine their own uniqueness and the unique nature of their succession. Every marker of the biochronological scale has its proper name and can be identified in all localities. Dissimilarly to paleontological markers, other markers such as paleomagnetic, lithological, tectonic, climatic, and others lack the distinctive individual features and cannot be adequately identified in different sections. Despite the probable shortness and even global character of corresponding geological events, they can hardly be used for creation of chronological scales without help of biochronological indicators. From this standpoint, chronological markers can be conditionally divided into autonomous (mainly paleontological indicators) and dependent (paleomagnetic data and some lithological levels, such as thin ash beds, tempestites, tidalites, eustatic facies, and others).

To analyze probable consequences of introducing dependent markers into zonal units in order to enhance the biochronological scale resolution, let us consider two sections significantly spaced apart, in which the same stratozones Z_2 are established based on the appearance levels of successive index species A_1, A_2, and A_3. The real position of species teilzones is estimated according to the conditional scale shown in the left part of the sketch (Fig. 4). Biozones of species can also be determined using the same scale. In the first section, teilzones of species are close to their biozones, and the stratozone spans the entire stratigraphic interval. In the second section, teilzones of these species are reduced, and the stratozone becomes reduced and slightly distorted so that its upper boundary turns to be within the overlying zone. In both sections, stratozones enclose thin bentonite beds, which are considered as certainly synchronous. Introducing this marker into the biochronological scale, we detaile it and obtain possibility to subdivide the zone into two parts: the lower one below tuffaceous bed and the upper one above the latter. As it is evident from the figure however, this new scale increases the correlation error for the lower part of the stratigraphic zone as compared with that characterizing simple correlation between the lower halves (by thickness) of stratozones in the first and second sections. There is no need to mention that the correlation accuracy reduces further, if we erroneously introduce the bentonite beds as additional markers to detail the previous scale, when they are different in age, though enclosed in the same stratozone (correlation between sections II and III in Fig. 4). It can be probably stated that introduction of dependent markers into chronological scales may preserve, never increases, and usually reduces the correlation accuracy. The reason is that, using the zonal biochronological scale, we cannot determine the real position of the intrazone marker in

the scale. In addition, synonymous stratozones in different sections are characterized by different volumes.

There were attempts to increase the resolution of biochronological scales with the help of zones distinguishable in sections based on some other fossil groups. However, introduction of new boundaries into zonal units, because they "split and diminish previously defined zones" (Gladenkov, 1995, p. 7), does no increase the correlation accuracy like in case with the bentonite marker. Moreover, the correlation uncertainty appears in this case because of changed compositional representation of zonal units in different sections, on the one hand, and because of variations in position of the newly introduced zone (based on other fossil group) within the stratozone of the initial scale, on the other. Figure 5 demonstrates reduction of the correlation accuracy of stratozones in the case under consideration.

Detailing of biochronological scales requires detailing of principal basis used to create the initial scale. When the species evolution was such a basis, we have nothing else to do, but to define the more detailed successive stages in their development. This is the way well known to paleontologists and stratigraphers who defined for a long time the subspecies, morphotypes, or morphological trends, in which evolutionary changes are detectable at the species and intraspecific levels.

The case when detailing concerns biostratigraphic scales, in which the boundaries of zonal units are facies-dependent, and the scales are used within a single sedimentary basin, needs a special remark. Introduction of additional lithological markers, which are of reference value in the study region, or of new paleontological levels that are also facies-dependent can result in a higher resolution of the scale, but only in the limited territories. I believe that all ways proposed by Gladenkov (1995) for the enhancement of scale resolution concern precisely this case and cannot be used for perfection of biochronological scales of a wider application.

CONCLUSION

(1) Principal features of stratigraphic scales are determined by the specifics of their basis. From this standpoint, two types of paleontologically substantiated stratigraphic scales can be defined: biochronological, based on the evolutionary succession of species (or formal taxons) and biostratigraphic, based on stratigraphic distribution of fossil assemblages. Position of zonal boundaries in the biochronological scale is strictly determined by the appearance level of index species. Zonal boundaries in the biostratigraphic scale are ambiguous, occurring within intervals between assemblages. Therefore, precise correlation of boundaries between stratigraphic units is possible only using biochronological scales.

(2) Zonal stratigraphic scales substantiated by paleontological data can be used only for correlation of stratigraphic units. Recognition of zones in sections (stratozones) represents an intermediate operation in application of the stratigraphic scale for correlation of stratigraphic units.

(3) The zone as elementary unit of the scale should be discriminated from the stratozone as a unit of the section. The latter is distinguishable as a result of applying a zonal scale to subdivision of the particular section, in which we conditionally correlate boundaries of zonal units in the scale with closest lithological boundaries. However, the scale does not inform us which part of the given zone is represented in the section. Accordingly, boundaries between stratozone subdivisions (upper, middle, lower), which are distinguishable based on lithological features, cannot be fixed in the paleontological scale to serve as the basis for the enhancement of its resolution.

(4) The continuity of the biochronological scale is ensured by using the evolutionary succession of fossil forms, distribution ranges of which are partly overlapped. The continuity of biostratigraphic scales is controlled by the continuity of successive fossil assemblages filling stratigraphic gaps that are discovered in the course of scale application.

(5) The chronological accuracy of correlation increases with the frequency increase of marker events in the homotaxial succession that is used as the basis for the chronological scale. In the practice, correlation can be performed with the maximum accuracy corresponding to the zone range.

(6) The enhanced resolution of biochronological scales can be achieved using the same means, which were used to create the initial scale, i.e., by the discovery of previously unknown members of the evolutionary succession of fossil forms. Intrazone markers, which differ in nature from those used to create a biochronological scale, may preserve, never increase, and usually reduce the correlation accuracy.

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Microfossils from the Riphean Hypostratotype of Siberia (the Omakhta, Kandyk, and Ust'-Kirba Microbiotas)

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Abstract—Organic-walled microfossils from the Riphean hypostratotype of Siberia (the Uchur–Maya region), which represent the Early Riphean Uchur (Omakhta) and Late Riphean Ui (Kandyk and Ust'-Kirba) microbiotas, are considered. The analyzed original materials (over 90 samples) show that the Omakhta microbiota is dominated by small to medium-sized *Leiosphaeridia, Myxococcoides, Eomycetopsis, Leiotrichoides, Oscillatoriopsis*, and other transit taxa (more than 20 genera and 30 species). New localities, which have been discovered, broaden our knowledge on the microfossil-bearing Omakhta deposits (18 microphytological levels in seven outcrops contrary to the five levels previously known at three sites). These levels are lacking large morphotypes of complex structure, which are present in other regions, the Early Riphean age of deposits is implied by isotopic-geochronological data. The microphytological characteristics of the Upper Riphean Kandyk–Ust'-Kirba interval of the hypostratotype is based on the analysis of 120 collected samples and published data. Microbiotas were found in eight microphytological levels of ten localities (versus six levels of eight sites known before). Forms dominant most representative microbiotas are *Chuaria, Simia, Valeria, Cucumiforma, Trachytrichoides, Polytrichoides, Arctacellularia, Palaeolyngbya*, and simple forms inherited from the underlying deposits. In addition, some levels of the Ui deposits contain *Trachyhystrichosphaera, Prolatoforma, Asperatofilum, Recta, Germinosphaera, Caudosphaera*, and some other morphotypes. Relationships between the Omakhta and Ui microbiotas and their comparison with other Riphean microfossil assemblages from Siberian and South Uralian type sections are considered.

Key words: Riphean, biostratigraphy, microfossils, microphytological level, terrigenous-clayey deposits.

The multidisciplinary substantiation of general and regional stratigraphic scales for the Upper Precambrian presumes the permanent improvement of main methods of subdivision and correlation of Proterozoic deposits (Semikhatov, 1995, 1997, and others). An important tendency of recent time is application of competitive geochronological (biochronometric) and biostratigraphic models for the ancient biosphere evolution.

According to the first one, ages of individual microbiotas are determined by different isotopic-geochronological methods of dating their host deposits (Semikhatov, 1991; Knoll, 1992, 1994, 1996; Knoll and Sergeev, 1995; Sergeev, 1997; Yakshin and Faizullin, 1997; and others). In this case, the general picture of the Riphean life evolution turns out to be closely dependent on dating results and represented by discrete microbiotas and even individual morphotypes, which are integrated into a single vertical succession without due regard to, or even in contempt of their morphological features (Xiao *et al.*, 1997). Microphytological practice shows that some complex taxa inevitably turn out to be transit throughout the Riphean in this case, and their first occurrence level becomes attributed conceptually to the Riphean base. Other microfossils find their position at few levels of taxonomic changes in the microbiota composition (the level of "Neoproterozoic revolution," for example). Age of these levels, like of phytolitic horizons actual before (Krylov, 1985), should be corrected after every significant revision of the chronometrical coordinates (see examples of the latest correction in Semikhatov *et al.*, 2000).

The alternative biostratigraphic concept of secular changes in the Late Proterozoic microbiotas is based on a thorough facies–ecological analysis of distribution of microfossil assemblages in ancient basins, on the study of representative regional successions of microfossils, and on the distinguished common features (homotaxy) of these successions (Veis, 1995, 1999). Ages of individual microbiotas in the successions are determined in this case by biostratigraphic correlation of their host deposits with the Riphean stratotype of the South Urals without a direct regard to isotopic-geochronological and chemostratigraphic data (Keller and Yankauskas, 1980; Yankauskas, 1982; Pyatiletov, 1988; Veis, 1988; Veis *et al.*, 1998a, 2000, 2001; and others). According to this concept, objects to be studied are the most complete Riphean successions, which must be comprehensively investigated in terms of microphytology to reveal main secular trends in the ancient microbiota changes under influence of evolutionary, paleoecological and taphonomic factors. Among the objects known to meet the requirements, the most important in Russia are

Riphean sections of the South Urals and Uchur–Maya region, and to some extent those of the Turukhansk and Anabar regions.

One of the primary objectives of the microphytological analysis of Riphean sections is to obtain information on taxonomic diversity and biological links of microorganisms, which populated basins the initial and terminal Riphean time. This is necessary to objectively evaluate changes in microbial communities close to the boundaries of this eon. The requirement of a rigorous solution of the problem decreases a number of objects suitable for investigation, because the very important Turukhansk section lacks the Lower and, probably, Middle Riphean beds (Khomentovskii *et al.*, 1985; Khomentovskii and Nagovitsin, 1998; Komar, 1990; Shenfil', 1991; and others), and the age interpretation of the Anabar section is highly disputable (compare: Gorokhov *et al.*, 1991, 1997; Pokrovskii and Vinogradov, 1991; Sergeev *et al.*, 1995; Veis and Petrov, 1994). Accordingly, the South Uralian stratotype and Uchur–Maya hypostratotype of the Riphean, especially their Early and Late Riphean microbiotas, are of prime significance at present for the comprehensive microphytological analysis.

In the Uchur–Maya region, investigations of this kind were neglected for more than ten years (Pyatiletov, 1988; Veis, 1988; Veis and Semikhatov, 1989). In this article, we intend to fill the gap with new data on more than 200 samples collected from the best sections of the Uchur and Ui groups, the lower and upper Riphean subdivisions in the region. The data are considered below with necessary references to published data. The used regional stratigraphic scheme of the Riphean is that substantiated by Semikhatov and Serebryakov (1983).

GEOLOGICAL POSITION OF THE OBJECTS STUDIED

The vast Uchur–Maya region extends over two significant tectonic structures of East Siberia: the southeastern margin of the Siberian platform and the adjacent western periphery of the Verkhoyansk–Kolyma foldbelt. Accordingly, in the region there are two different areas of distribution of the Riphean deposits: the stable Uchur–Maya plate in the southwest and the Yudoma–Maya belt of rifting origin in the east (Fig. 1). The Omnya uplift in the Uchur–Maya interfluve divides the plate into the Uchur and Maya depressions. The submeridional Yudoma–Maya belt includes a series of subordinate structural elements (Nuzhnov, 1967; Pushcharovskii, 1972; Semikhatov and Serebryakov, 1983; Shenfil', 1991). The Upper Precambrian sections of the Uchur–Maya plate and Yudoma–Maya belt differ in completeness, depositional environment, thickness, and mode of occurrence. The Riphean and discordantly overlying Yudomian beds are subhorizontal in the plate and deformed into box folds, faulted parallel to the structural axis, in the Yudoma–Maya belt.

The general Upper Precambrian succession of the Uchur–Maya region includes six large sedimentary megacycles corresponding to the Uchur, Aimchan, Kerpyl', Lakhanda, Ui, and Yudoma groups. Five lower groups are referred to the Riphean and the sixth one to the Vendian. As mentioned above, we consider here microfossils of the Uchur and Ui groups, one beginning and the other terminating the Riphean succession in the region.

For a long period, the isotopic-geochronological dates characterizing Riphean deposits of the Uchur–Maya region were those obtained by K–Ar method for mineralogically unstudied glauconite in the 1960s–1970s. They represented successions of age values descending from 1500 to1360 Ma in the Uchur Group and from 760 to 700 Ma in the Ui Group (Semikhatov and Serebryakov, 1983). Recent results made these age intervals noticeably shorter. The U–Pb dates for volcano-plutonic complex that underlies the Uchur Group and constrains its maximum age range from 1703 ± 18 to 1727 ± 11 Ma (Larin *et al.*, 1997). Basic sills that crosscut the Ui Group yielded the U–Pb ages of 974 ± 1 and 1005 ± 4 Ma, and the Sm–Nd age of 942 ± 18 Ma (Pavlov *et al.*, 1992; Rainbird *et al.*, 1998). Owing to these and other dates obtained recently, estimation of the isotopic age of an important chronostratigraphic boundary in the Riphean (the lower limit of the Karatavian Erathem) increased up to 1030 Ma (Semikhatov *et al.*, 2000), and this correction allowed recognizing of analogues of three main Riphean subdivisions in the region. On the other hand, the K–Ar and Rb–Sr dates for whole-rock samples of different-aged clayey deposits (including those the Uchur and Ui groups) served as a ground to suggest that the entire Riphean succession in the Uchur–Maya region may correspond to the Middle Riphean erathem only (Vinogradov, 1995; Vinogradov *et al.*, 2000).

Deposits of the Uchur Group widespread in the synonymous depression are also exposed in a small area of the northern Yudoma–Maya belt. In the former structure, the distinctly transgressive deposits are subdivided into three units: the predominantly sandstone Gonam, rhythmical carbonate-siliciclastic Omakhta, and clayey-dolomitic Ennin formations, the latter including sandstone horizons (these units are 185–200, 170–180, and 200–220 m thick, respectively). In the northern Yudoma–Maya belt, the central part of the Gornostakhskii anticline is composed of the siliciclastic-carbonate Trekhgorka (1100–1200 m) and Dim (1800–2000 m) formations, which are commonly considered as stratigraphic equivalents of the Uchur Group. We investigated the entire distribution area of the Uchur deposits along the Belaya, Idyum, Algoma, Bol'shoi Aim, and Uchur rivers (Fig. 1).

The Riphean succession of the region is terminated by the thick siliciclastic Ui Group consisting of two regional units: the predominantly sandstone Kandyk and siltstone–mudstone Ust'-Kirba formations, which

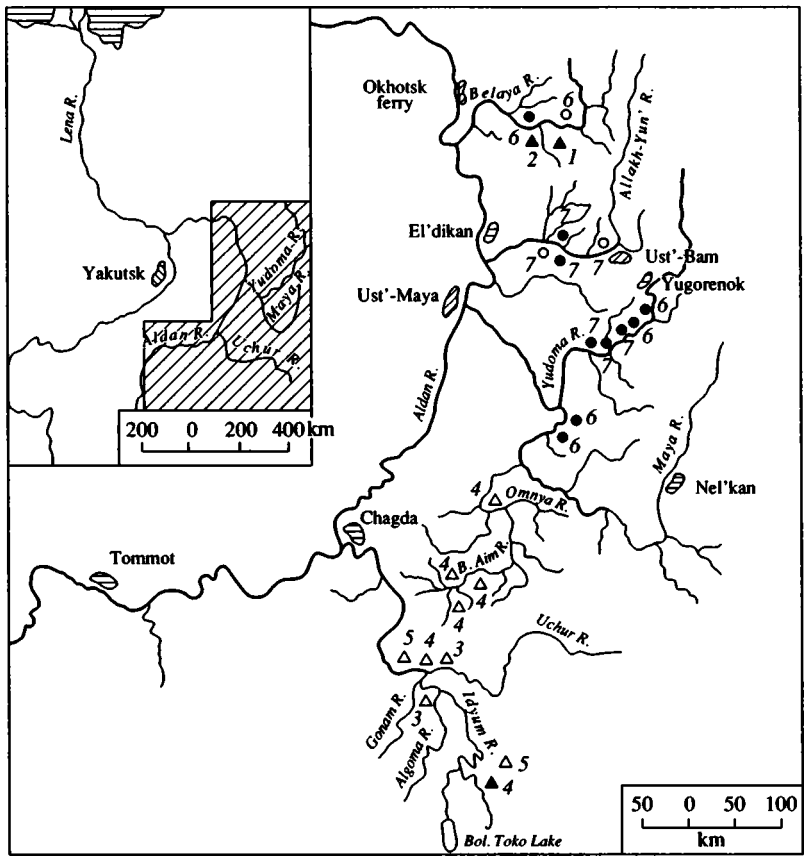


Fig. 1. Sampling sites of the Uchur and Ui groups in the Uchur–Maya region of Eastern Siberia. Formations of the Uchur Group (triangles): *1*, Trekhgorka, *2*, Dim, *3*, Gonam, *4*, Omakhta, *5*, Ennin. Formations of the Ui Group (circles): *6*, Kandyk, *7*, Ust'-Kirba. Localities of samples with and without microfossils are designated by solid and open symbols, respectively.

predominantly occur in the Yudoma–Maya belt, where they are 3000 and 1500 m thick. Outside, they are known only in the eastern margin of the Maya depression, where the sandstones and siltstones of the Kandyk Formation about 220 m thick survived the pre-Yudomian erosion. Rock varieties of the Ui Group, which are most suitable for microphytological analysis, were sampled along the Belaya, Allakh-Yun', Yudoma, and Maya rivers (Fig. 1).

Prior to discussion of data obtained, we should note that more than 200 samples were treatment in laboratory. A half of them yielded scarce to abundant microfossils of a good to satisfactory preservation degree. The classification of organic-walled microfossils from the Uchur and Ui groups and methods of their extraction from deposits do not differ from those described earlier (Veis and Vorob'eva, 1992, 1993). The taxonomic composition of studied microbiotas is considered below.

MICROFOSSILS OF THE UCHUR GROUP

Up to the present, microfossils of this group were thoroughly studied only from three localities of the Omakhta Formation in the southern Uchur depression, where the rocks are exposed along the Idyum River (Veis and Semikhatov, 1989). In addition, occurrences of single organic-walled sphaeromorphids, represented by discarded taxa from the Uchur Group (Timofeev, 1966, 1969), and presence of rare coccoidal *Myxococcoides solitaria* Kol., *Dzhandophyton globosus* Kol., and *D. formosus* Kol. in the Omakhta cherts at the Uchur River (Kolosov, 1982) were reported. However, the validity of two latter taxa is disputable (*Mikrofossilii...*, 1989), and the first of them is most likely a younger synonym of the simplest *Myxococcoides* form distinguished earlier. In 1979–1993, our study in the northern part of the Uchur depression, along the Uchur, Algoma, Aim, and Bol'shoi Aim rivers (Fig. 1), revealed scarce microfossils confined to the Omakhta Formation in the coastal shallow-water deposits of the Uchur Group (Semikhatov and Serebryakov, 1983). Fifty four Omakhta samples yielded single medium-sized *Leiosphaeridia* only, whereas 20 samples of the Gonam and Ennin formations turned out to be barren. Investigation of the Omakhta deposits in the southern part of the depression was more fruitful in the microphytological respect. Fairly representative microbiotas were found in deposits of the relatively deep-water set-

tings remote from the coast and isolated from the intense supply of clastic material.

In the other distribution area of Lower Riphean deposits (the northern part of the Yudoma–Maya belt), the microphytological analysis was carried our for deposits of the Trekhgorka Formation, which are exposed along the Svetlyi Creek crossing the central Gornostakhskii anticline (four of 18 samples yielded microfossils), and for rocks of the Dim Formation sampled along the Trekhgorka (two of 14 samples contained microfossils) and Belaya river valleys (five of 12 samples were characterized by microfossils). Both formations enclose taxonomically similar assemblages of low diversity, which include rare and small dark-colored *Leiosphaeridia* (up to 60 μm), single *Satka* (up to 50 μm), colonial coccoidal *Myxococcoides* (cell diameter 8–10 μm), and filamentous sheaths of *Eomycetopsis* and *Leiotrichoides* (from 3–4 to 10–14 μm wide).

The characteristic composition of the diverse Omakhta microbiota is presented below. It was found in one of the best exposure of the synonymous formation in the Idyum River middle courses upstream and downstream of the Malyi Sivaglikan Creek mouth, the southern part of the Uchur depression. The rhythmical Omakhta Formation consists of three units (Semikhatov and Serebryakov, 1983). Our investigation noticeably improved the previous knowledge of taxonomic composition of microfossils in each unit (Veis and Semikhatov, 1989). The complete taxonomic composition of found microfossils and their stratigraphic positions in the studied sections are shown in Fig. 2, and characteristic morphotypes are pictured in Plates I and II. The distinguished 18 microphytological levels are numbered from the base upward and characterize an integral succession of the Omakhta deposits.

The first unit with apparent thickness of 130–140 m (1st to 6th microphytological levels) is most completely exposed at the Idyum River, within the interval of 6–15 km downstream the Malyi Sivaglikan Creek mouth. Earlier, only one exposure located at the right bank of the Idyum River, 9.5 km downstream of the creek, was thoroughly studied. Five of 12 samples from the 16-m-thick interval of the lower part of the unit yielded microfossils belonging to six genera and six species (Veis and Semikhatov, 1989). In this work, we studied 22 samples. Microfossils were found in five samples collected at six discrete levels of three exposures at the right bank of the Idyum River (14, 9.5, and 6 km downstream the Malyi Sivaglikan Creek mouth). Regretfully, there are few dark-colored clayey beds suitable for the microphytological analysis, and newly obtained data did not improve much the previous paleontological characteristics of the unit. Microfossils of the levels are described below.

The first and second levels (the right bank of the Idyum River, 14 km downstream the Malyi Sivaglikan Creek mouth; one sample with microfossils from each level) yield rare thick-walled, heavily corroded *Leiosphaeridia* (30–70 μm in diameter) and small shapeless colonies of *Myxococcoides* (cell size 8–10 μm).

The third to fifth levels (the right bank of the Idyum River, 9.5 km downstream the Malyi Sivaglikan Creek mouth; two samples with microfossils from each level) enclose a more diverse assemblage. Previously found forms were rare *Leiosphaeridia* and *Nucellosphaeridium* (10–55 μm in diameter), and single *Oscillatoriopsis, Cyanonema*, and *Gloeodiniopsis*. The newly found forms are *Leiosphaeridia* of larger size (up to 90 μm), accumulations of envelopes each up to 40 μm wide, and rare oscillatorian trichomes (6–8 μm wide). Missing representatives of *Cyanonema* and *Gloeodiniopsis* are replaced by colonial coccoidal *Myxococcoides* (cell size 12–14 μm) and by sheaths of *Eomycetopsis* and *Leiotrichoides* (3 or 4 and 8–14 μm, respectively).

The six level (the right bank of the Idyum River, 6 km downstream of the Malyi Sivaglikan Creek mouth; one sample with microfossils) reveals occurrence of scarce small to medium-sized acritarchs *Leiominuscula* (8–10 μm), *Leiosphaeridia*, and *Nucellosphaeridium* (30 and 70 μm in diameter), colonial coccoidal *Myxococcoides* (cell size 14–16 μm), and interwoven *Eomycetopsis* sheaths (2–3 μm wide).

Plate I. Microfossils from the Omakhta Formations of the Riphean Uchur Group, the Uchur–Maya region.

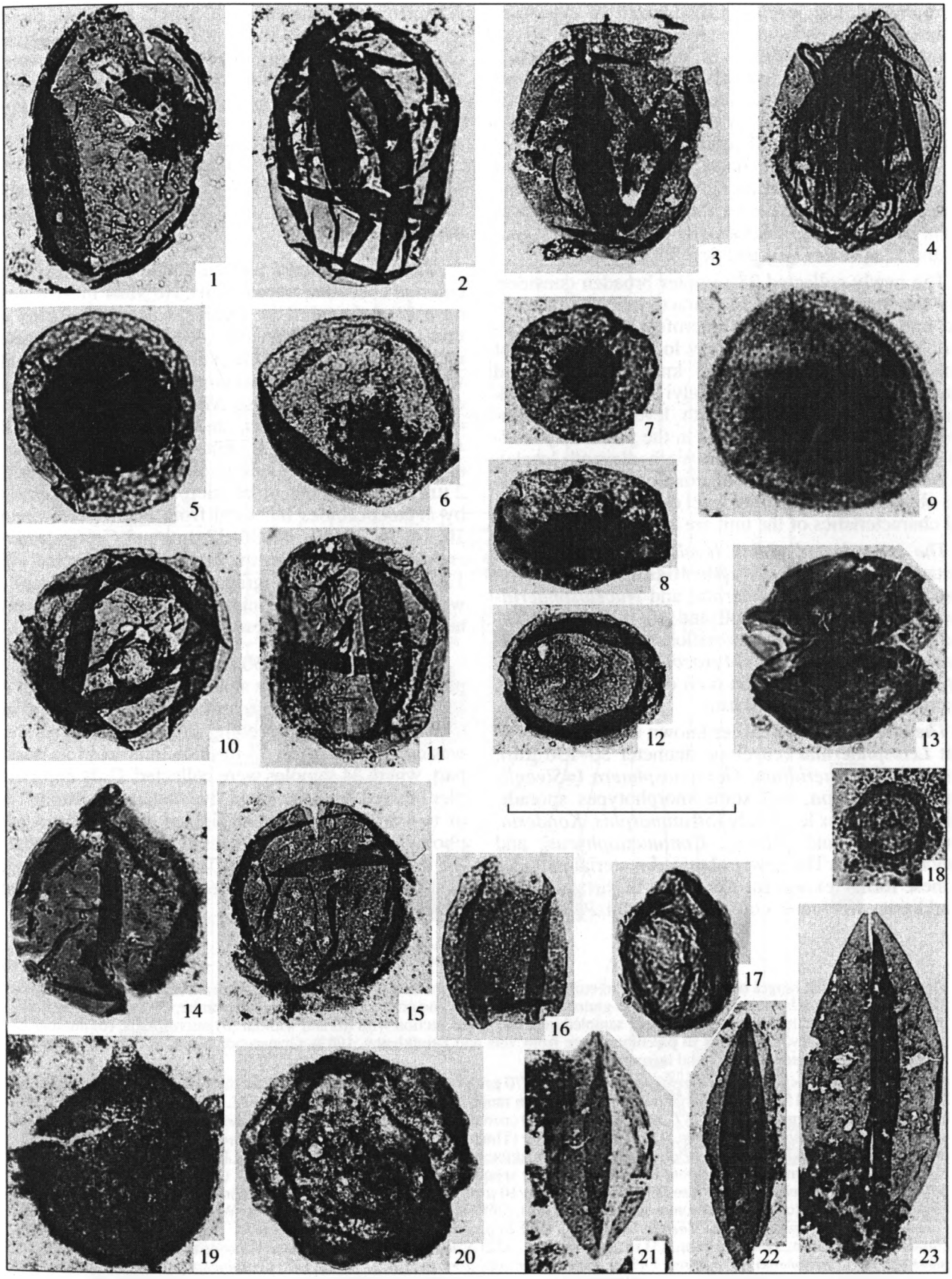
All specimens pictured here are stored at Geological Institute of Russian Academy of Sciences, Moscow.

(8, 13) eleventh microphytological level, second unit, the right bank of the Idyum River, 0.5 km downstream of the Malyi Sivaglikan Creek mouth; (5, 7, 11, 16, 17) thirteenth microphytological level; (1–4, 9, 10, 12, 14, 15, 18–23) fourteenth microphytological level; (6) seventeenth microphytological level, third unit; the right bank of the Idyum River, 4 km upstream of the Malyi Sivaglikan Creek mouth.

Scale bars: 120 μm for (1, 4, 9, 10, 12, 20); 100 μm for (2, 3, 6–8, 11, 15); 80 μm for (5, 13, 14, 16, 18, 19); 60 μm for (17); 180 μm for (21); 220 μm for (22); and 160 μm for (23).

(1–4, 10–12, 14, 15) *Leiosphaeridia tenuissima* Eisenack: (1) sp. no. 933/1319, (2) sp. no. 933/1414, (3) sp. no. 933/1311, (4) sp. no. 933/1306, (10) sp. no. 933/1310, (11) sp. no. 933/1301, (12) sp. no. 933/1301, (14) sp. no. 933/1304, (15) sp. no. 933/1317; (5) *Pterospermopsimorpha pileiformis* Tim., emend. Mikh., sp. no. 933/4178; (6) *Leiosphaeridia incrassatula* Jank., sp. no. 933/2119; (7) *Nucellosphaeridium minutum* Tim., sp. no. 933/4152; (8) *Konderia* sp., sp. no. 934/5144; (9) *Pterospermopsimorpha granulata* Mikh., sp. no. 933/1510; (13) *Leiosphaeridia bicrura* Jank., sp. no. 934/5215; (16, 17) *L. crassa* (Naum.), emend. Jank.: (16) sp. no. 933/4110b, (17) sp. no. 933/4125; (18) *Simia* sp., sp. no. 933/7223; (19) *Germinosphaera tadasii* A. Weiss, sp. no. 933/1529; (20) *Leiosphaeridia laminarita* (Tim.), emend. Mikh., sp. no. 933/1318; (21–23) *L. jacutica* (Tim.), emend. Mikh. et Jank.: (21) sp. no. 933/1411, (22) sp. no. 933/7270, (23) sp. no. 933/1428.

Plate I



The second unit 160 to 180 m thick (7th to 11th microphytological levels) is recognized in exposures complementing each other along the Idyum River, downstream and upstream of the Malyi Sivaglikan Creek mouth. Earlier, we studied in detail the 13-m-thick upper fragment of the unit. The fragment is exposed at the left bank of the Idyum River, 1.8 km upstream of the Malyi Sivaglikan Creek mouth. Only two of eight samples contained microfossils. These two samples were collected along the strike of dolomitic mudstone bed from the level 2.4–2.6 m above the base of the fragment. The microfossils belonged to 11 genera and 12 species (Veis and Semikhatov, 1989).

The newly collected 28 samples broaden considerably the microphytological characteristics of the second unit (14 samples with microfossils). We discovered, on the one hand, two new localities at the right bank of the Idyum River, 0.5 km downstream and immediately upstream the Malyi Sivaglikan Creek mouth (the seventh and eighth levels), and distinguished additional microfossils in the previously studied and overlying deposits (the ninth–eleventh levels), on the other. The previous information on microfossils was obtained from the ninth level only. Microphytological characteristics of the unit are as follows.

The seventh and eighth levels yields innumerous acritarchs *Leiominuscula* (up to 10 μm in diameter) in association with *Leiosphaeridia* and *Nucelosphaeridium* (envelope sizes 30–160 and 40–60 μm, respectively). The eighth level also encloses isolated accumulations of small coccoidal *Myxococcoides* sp. of peculiar elongated configuration (cell diameter 7 or 8 μm, accumulation size 40 × 90 μm).

The ninth level was earlier known to contain dominant *Leiosphaeridia* (envelope diameter 30–130 μm), rare *Nucellosphaeridium, Germinosphaera* (=*Sivaglicania*), *Polycavita*, and some morphotypes sporadically found at this level only (*Stratimorphis, Konderia*, small *Simia* and *Aimia*, *Coniunctiophycus* and *Siphonophycus*). The newly obtained material includes all these forms (except for *Konderia*) of the same size. In addition, we found colonial coccoidal *Palaeoanacystis* and *Myxococcoides*. The former occurs in dense subspherical aggregates (cell size 5 to 6 μm, dimensions of aggregates 100 × 120 μm), and the latter is clustered into small (up to 150 μm) shapeless accumulations of cells 7–8, 10–12, and 20–22 μm in diameter. Rare narrow *Eomycetopsis* sheaths (4 μm wide) supplement the assemblage.

The tenth level encloses only innumerous *Leiosphaeridia* (20–80 μm in diameter) represented by thick-walled smallest and thin-walled larger specimens.

The eleventh level yielded abundant acritarchs and colonial coccoidal microfossils. No other morphotypes have been found. The dominant acritarch forms are thin-walled *Leiosphaeridia* (40–120 μm in diameter) including *L. bicrura* Jank. and *L. incrassatula* Jank., which frequently form accumulations of few and more envelopes. There are also *Nucellosphaeridium* (up to 40 μm), dividing forms, and small envelopes (30–40 μm) of unclear taxonomic affinity, which occur as aggregates of closely spaced dark-colored granules (1–2 μm). Colonial coccoidal microfossils are represented by *Myxococcoides* inherited from below (cell size 8–10, 14–16, 18–20, and 30–32 μm) and *Coniunctiophycus* (cell size 10–20 μm, diameter of colonies 100–120 μm, and that of aggregates up to 200–220 μm), as well as palmelloid colonies of *Eoentophysalis* (diameter of *capsulata* and *capsulopunctata* cells 3–4 μm).

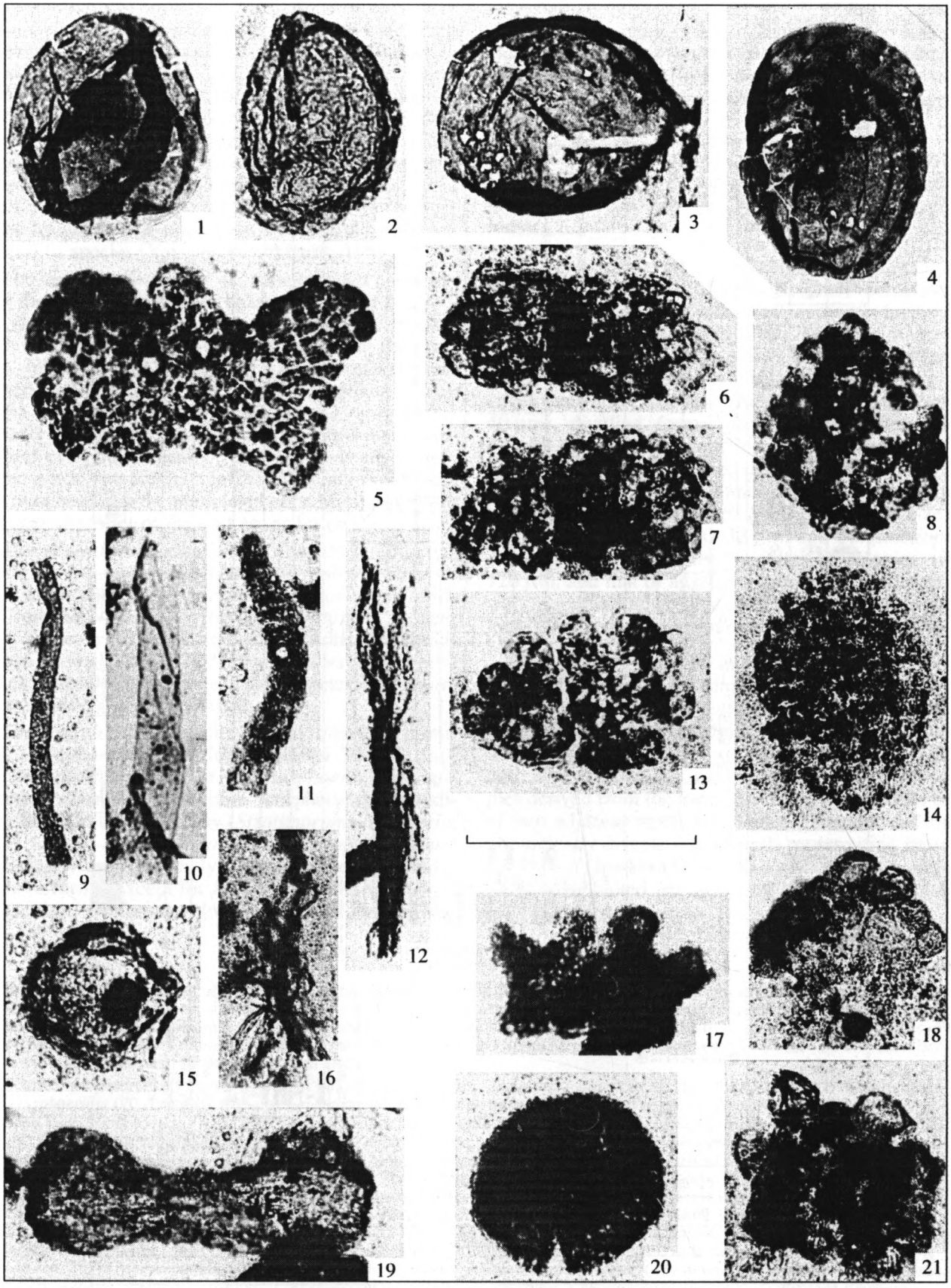
The third unit 48–50 m thick (12th to 18th microphytological levels) was studied at the right bank of the Idyum River, 4 km upstream of the Malyi Sivaglikan Creek mouth. The preceding microphytological study was carried out in the 13-m-thick interval of its middle part, where 34 samples were collected. Only eight samples derived from the lower five meters (mudstone beds of two adjacent rhythms) yielded abundant and morphologically diverse microfossils representing most of the known Omakhta forms. The microfossils belonged to 18 genera and 26 species (Veis and Semikhatov, 1989).

Fig. 2. Stratigraphic ranges of organic-walled microfossils from the Omakhta Formation of the Riphean Uchur Group, the Uchur–Maya region: (1) mudstone, siltstone, and fine-grained sandstone, (2) fine grained and fine-clastic dolomite, (3) stromatolitic dolomite, (4) coarse-clastic dolomite; (5) closely sampled intervals of the section with (*a*) and without (*b*) microfossils; (6) number of microphytological level (numbers in parentheses are from Veis and Semikhatov, 1989). Occurrence levels of rare and abundant microfossils is indicated by small and larger circles.

Pictured morphotypes: (1) small *Leiosphaeridia* (less than 70 μm), including *Leiominuscula* (less than 10 μm), (2) medium-sized *Leiosphaeridia* (70–150 μm), (3) large *Leiosphaeridia* (size range from 150 to 230–240 μm), (4) *L. incrassatula* Jank., (5) *L. bicrura* Jank., (6) *L. ternata* (Tim.), (7) *L. kulgunica* Jank., (8) coenobial aggregates of small *Leiosphaeridia*, (9) *Nucellosphaeridium* spp., (10) *Pterospermopsimorpha* spp., (11) *Simia annulare* (Tim.), emend. Mikh., (12) *Simia nerjenica* A. Weiss, (13) *Satka granulosa* Jank., (14) *Konderia* sp., (15) *Polycavita* sp., (16) *Stratimorphis* sp., (17) *Eomarginata stricta* Jank., (18) *E. idumii* A. Weiss, (19) *Aimia* sp., (20) envelopes composed of rows of closely spaced granules (granule size 1–2 μm), (21) *Palaeoanacystis vulgaris* Schopf, (22) *Myxococcoides* (cell size in colonies less than 10 μm); (23) *Myxococcoides* (cell size in colonies from 10 to 20 μm), (24) *Myxococcoides* (cell size in colonies more than 20 μm), (25) *Coniunctiophycus omachticus* A. Weiss, (26) *Gloeodiniopsis* sp., (27) *Palaeopleurocapsa* sp., (28) *Eoentophysalis* sp., (29) *Eomycetopsis robusta* Schopf, emend. Knoll, Golubic, (30) *Leiotrichoides typicus* Herm., emend. Herm., (31) *L. gracilis* Pjat., (32) *Siphonophycus* sp., (33) *Oscillatoriopsis* spp., (34) *Filiconstrictosus* sp., (35) *Cyanonema* aff. *inflatum* Oehler, (36) *Germinosphaera tadasii* A. Weiss, (37) *Germinosphaera* with characteristic "cerebrospheric" texture of envelope, (38) supposed sporangia-like structures, (39) dumbbell-like forms of complex structure.

Group
Formation
Unit
Thickness, m
Lithological column
Sampled interval
Microphyto-logical level
Microfossils
Uchur
En
Omakhta
Gn
Third
Second
First
48–50
160–180
130–140
The Idyum R., 14; 9.5; and 6 km downstream the Malyi Sivaglikan Creek mouth
The Idyum R., near the Malyi Sivaglikan Creek mouth
The Idyum R., 4 km upstream the Malyi Sivaglikan Creek mouth

Plate II



New materials are obtained from 49 samples collected in the same exposure at the Idyum River. The 25-m-thick interval that was sampled includes the aforementioned 13-m-thick sequence that was extended 2 m downward (owing to the lower water stand) and 10 m upward. Spacing between sampling sites was relatively regular. Twenty samples of clayey-carbonate light-colored interbeds unsuitable for microphytological analysis contain no identifiable forms. The other 29 samples of dark-colored mudstones yielded variably abundant microfossils, which are confined to seven microphytological levels. The second and third levels (the thirteen and fourteen levels in the general succession) reveal a complete paleontological characteristics of the third unit, which are as follows (Fig. 2).

The twelfth level encloses only rare thin- and thick-walled acritarchs *Leiosphaeridia* (envelope diameter 70–90 μm), among them specimens of *L. incrassatula* Jank. with tylosislike thickening and those of *L. bicrura* Jank. with a median fissure.

The thirteen level contains a more diverse microfossil assemblage. Numerous acritarchs (up to 150 μm in diameter) and less abundant colonial coccoidal and filamentous microfossils were previously reported from this level. They all represent 11 genera and 19 species. The transit forms *Leiosphaeridia, Nucellosphaeridium, Pterospermopsimorpha, Myxococcoides, Leiotrichoides*, and *Eomycetopsis*, which are common in the Lower Riphean, were encountered together with rare *Germinosphaera* (=*Sivaglicania*), *Satka*, and *Palaeopleurocapsa*, and with evidently allochthonous *Eoentophysalis* almost unknown in clayey deposits. In the newly obtained material, all the above morphotypes associate with frequent thin- and thick-walled *Leiosphaeridia* (diameter up to 180 μm; the largest size 70–90 and 110–140 μm), marginated *Simia* of two size generations (30 and 80–100 μm), and single poorly transparent tubercular envelopes assigned to *Polycavita* (100 μm in diameter).

The fourteenth level is most rich in microfossils among the others. According to earlier data, its assemblage includes 18 genera and 26 species of different morphological structure. The main forms are transit *Leiosphaeridia* (10–150 μm in diameter) dominating in all samples, *Nucellosphaeridium, Pterospermopsimorpha*, and all colonial coccoidal and filamentous forms (sheaths and trichomes) of the preceding level. The high taxonomic diversity of the level is defined by presence of *Palaeoanacystis, Eomarginata, Polycavita, Filiconstrictosus, Oscillatoriopsis*, and other taxa, which associate with single complicated, supposedly sporangium-like forms found nowhere else in the Omakhta deposits. *Leiosphaeridia* of simple structure dominate in new material as well, but their maximal diameter is 240 μm. These larger forms are scaphoid, having longitudinal fold and thus correspond to diagnosis of the rejected genus *Macroptycha* Tim. Smaller specimens are of subspherical shape. Forms formerly unknown at this level are peculiar *Germinosphaera* with a single large process and characteristic cerebriform folds (80 μm in size), sheaths of *Siphonophycus* with indistinct transverse ornamentation (28–30 μm wide), peculiar dumbbell-like forms (80 × 140 μm), and tubercular *Polycavita* (60 and 80 μm in diameter), the latter inherited from the underlying deposits.

The fifteenth level encloses scarce *Leiosphaeridia* (envelope diameter 50–100 μm), some of which are fossilized at the division stage, and infrequent *Eomycetopsis* (2 or 3 μm) and *Leiotrichoides* (12–14 and 20–22 μm). The largest sheaths are represented by their fragments, and the narrowest ones are locally interwoven.

The sixteenth–eighteenth levels contain rare *Nucellosphaeridium* and *Leiosphaeridia* (diameter 40–110 and 80–90 μm, respectively). Envelopes of *L. incrassatula* Jank. with tylosislike thickening were found at the seventeenth level, and accumulations of small *Leiosphaeridia* (each envelope 40 μm in diameter), *Myxococcoides* (cell size 7 to 8 μm), and single *Leiotrichoides* (12–14 μm wide) occur at the eighteenth level.

Thus, the much more representative material (52 samples with microfossils from 18 levels of seven sites versus 15 samples from six levels of three sites)

Plate II. Microfossils from the Omakhta Formation of the Lower Riphean Uchur Group, the Uchur–Maya region.

All specimens pictured here are stored at the Geological Institute of Russian Academy of Sciences, Moscow.

(14, 20) ninth microphytological level; (5–9, 12, 13, 15, 17, 18, 21) eleventh microphytological level, second unit, the right bank of the Idyum River, 0.5 km downstream the Malyi Sivaglikan Creek mouth; (2) thirteenth microphytological level; (1, 3, 4, 10, 11, 16, 19) fourteenth microphytological level, third unit, the right bank of the Idyum River, 4 km upstream the Malyi Sivaglikan Creek mouth.

Scale bars: 120 μm for (1, 3, 4); 80 μm for (2, 10, 11, 20); 60 μm for (5, 21); 160 μm for (6, 8, 13); 200 μm for (7, 17); 50 μm for (9, 12, 16); 140 μm for (14, 18); 100 μm for (15); 70 μm and for (19).

(1–4) *Leiosphaeridia tenuissima* Eisenack: (1) sp. no. 933/1328, (2) sp. no. 933/4111, (3) sp. no. 933/1418, (4) sp. no. 933/1421; (5) *Eoentophysalis belcherensis* Hofmann, sp. no. 934/5248; (6, 7, 13) *Coniunctiophycus omachticus* A. Weiss: (6) sp. no. 934/5211, (7) sp. no. 934/5255, (13) sp. no. 934/5284; (8, 14, 18, 21) *Myxococcoides* spp.: (8) sp. no. 934/5266, (14) sp. no. 934/3274, (18) sp. no. 934/6329, (21) sp. no. 934/5256; (9, 12, 16) *Eomycetopsis robusta* Schopf. emend. Knoll, Golubic: (9) sp. no. 934/5261, (12) sublinear aggregate of sheaths, sp. no. 934/6308, (16) a fragment of interwoven sheaths, sp. no. 933/1520; (10) *Leiotrichoides typicus* Herm., emend. Herm., sp. no. 933/1416; (11) *Oscillatoriopsis media* Mendelson et Schopf, sp. no. 933/1314; (15) *Nucellosphaeridium minutum* Tim., sp. no. 934/5237; (17) *Coniunctiophycus* sp., sp. no. 933/1312; (19) dumbbell-shaped structure, sp. no. 933/1312; (20) *Polycavita* sp., sp. no. 934/3261.

did not change significantly the former data concerning taxonomic composition of the Omakhta microbiota. This fact is of great important, because this microbiota determines microphytological characteristics of the Uchur Group corresponding to the synonymous interregional horizon of Siberia (Khomentovskii *et al.*, 1985; Shenfil', 1991; Khomentovskii, 1996; Yakshin and Faizullin, 1997; and others). Our study specified distribution of some transit forms and allowed us to discover *Simia annulare* (Tim.) emend. Mikh. Other changes in the species composition of microbiota that was established earlier (Veis and Semikhatov, 1989) are as follows. Small structureless envelopes, formerly classed erroneously with *Protosphaeridium*, are now determined as *Leiosphaeridia* sp. *Sivaglicania tadassii* A. Weiss and *Chuaria* aff. *nerjenica* A. Weiss are attributed to genera *Germinosphaera* and *Simia*, respectively (*Mikrofossilii...*, 1989). At present, 33 species representing 24 genera define the Omakhta microbiota composition.

The different abundance and diversity of microfossil at the distinguished microphytological levels (Fig. 2) is most likely controlled by facies–ecological and, partly, by taphonomic factors (Veis and Semikhatov, 1989). It should be noted that the Omakhta microfossils are found in the middle horizons of thin (1–3 m) transgressive rhythms. The lower members of the rhythms are composed of sandstone showing signs of frequent subaerial exposition. Their middle members are composed of siltstone and mudstone, some of which are dolomitic and exhibit desiccation cracks disappearing upward the rhythm section. The upper members consist of dolomites, oolitic and/or sandy varieties of which are replaced upsection by stromatolitic varieties and, locally, by thin-bedded dolomicrites (Semikhatov and Serebryakov, 1978, 1983). The dolomitic members of the rhythms are interpreted as sediments of intertidal and upper subtidal zones, whereas the underlying clayey and carbonate-clayey members are attributed to sediments of protected settings the upper subtidal and/or lower intertidal zones (Semikhatov and Serebryakov, 1979; Veis and Semikhatov, 1989). The inferred sedimentation settings of carbonate members of the transgressive rhythms are confirmed by data on genesis of the Omakhta stromatolitic microstructures (Knoll and Semikhatov, 1998).

Thus, the comparison of the newly obtained and previously known materials shows the size increase of the Omakhta acritarchs *Leiosphaeridia* from 150–155 to 230–240 μm, which is remarkable in biostratigraphic aspect, and the parallel growth in abundance of contrastingly very small *Simia* (about 30 μm). Among the other morphotypes, there are distinguished formerly unknown *Germinosphaera* with the characteristic cerebriform structure, dark-colored tubercular forms (100 μm across) that do not belong to *Polycavita*, clustered *Leiosphaeridia* envelopes of possible coenobial nature, interwoven filamentous *Eomycetopsis*, and unnamed dumbbell-like forms of complicated structure (80 × 140 μm). Most of the Omakhta microfossils listed above are found only at the thirteenth and fourteenth levels of the third unit, in the mudstone interbeds accumulated under calm hydrodynamic environments far from the shore line of the Omakhta basin.

MICROFOSSILS OF THE UI GROUP

Siliciclastic deposits of this group are partially represented by dark-colored clay–siltstone interbeds suitable for accumulation of organic-walled microfossils. These deposits, yielding variably abundant microfossils, were studied by many researchers (Timofeev, 1966, 1969; Timofeev *et al.*, 1976; Rudavskaya and Frolov, 1974; Khomentovskii and Pyatiletov, 1978; Volkova *et al.*, 1980; Volkova, 1981; Pyatiletov, 1988; Veis, 1988; *Mikrofossilii...*, 1989; and others). The geographically discrete position of main microfossil localities, distinguished in the Ui Group within the vast Uchur–Maya region, and formerly published unsystematic data, characterizing stratigraphic position of microbiotas in the facies-variable deposits of the group, emphasize the actuality of new microphytological analysis in this interval of the Riphean hypostratotype of Siberia.

In this work, we accept the traditional subdivision of the Ui Group into the regional Kandyk and Ust'-Kirba formations (Semikhatov and Serebryakov, 1983). Many Siberian geologists distinguish instead three local subdivisions in the southern sections of the group, namely the Kandyk, Dzhabatyma (the Kandyk Formation *s.l.* is their equivalent) and Ust'-Kirba formations, or they subdivide the group into four units in the northern sections, where these are known as the Kandyk, Malaya Sakhara, Dzhoron (the traditional interval of the Kandyk Formation), and Dalynda or Ust'-Kirba, formations (Basharin and Samozvantsev, 1964; Volkodav *et al.*, 1978; Pyatiletov, 1988; *Resheniya...*, 1983; and others). These local subdivisions of the Ui Group, when they are mentioned below, are used for a more accurate positioning of microphytological levels, which we consider in a single succession. Because the published data vaguely characterize stratigraphic position of microfossils in the main sections of the Ui Group, these levels (in contrast to those of the Omakhta Formation) may span considerable intervals of the section (up to tens of meters thick). Below we discuss former and newly obtained data on taxonomic composition of microfossils from the Ui Group. Figure 3 illustrates the complete composition of microbiota and stratigraphic position of microphytological levels. Plates III and IV show the most characteristic forms.

The Kandyk Formation yields microfossils from five microphytological levels. The first level corresponds to the transitional interval (about 10 m thick) from upper carbonate beds of the Lakhanda Group to clay–siltstone beds of Kandyk Formation in the stratotype of the latter situated in the northern Maya depression, near the mouth of the Bol'shoi Kandyk River, the

right tributary of the Maya River. The second level is established far away from the first locality, in the Yudoma–Maya belt. It corresponds to the lower part of the Kandyk Formation (18–20 m thick) in the section that is located in the western wing of the Gornostakhskii anticline (exposures at the Belaya River near the Petrov-Yurege Creek mouth) and was partly eroded in the pre-Yudomian time. The third level is recognized in the upper part (approximately 10–20 m thick) of the Kandyk Formation *s. str.* that is exposed at the Keatan-Ytyga River (the Ryabinovyi Creek, Yudoma River basin) the central part of the belt. The fourth level more than 30 m thick corresponds to the overlying beds of the Malaya Sakhara and Dzhoron formations of the same exposure. The fifth level 10–14 m thick is distinguished in the geographic proximity to locality of the first level, at the top of the Kandyk Formation fragmentary exposed at the right bank of the Maya River, 4 to 5 km downstream the Bol'shoi Kandyk River mouth.

Microfossils of the first level were thoroughly studied by many researchers (see references in Pyatiletov, 1988). Data on second and fifth levels were published by Veis (1988) and obtained in this study. The third and fourth levels were characterized by Pyatiletov (1988). The new materials are obtained from 54 samples characterizing the first, second and fifth levels (12, 20, and 22 samples, respectively). Forty four of these samples (12, 10 and 22 from levels listed above) yielded abundant and well-preserved microfossils.

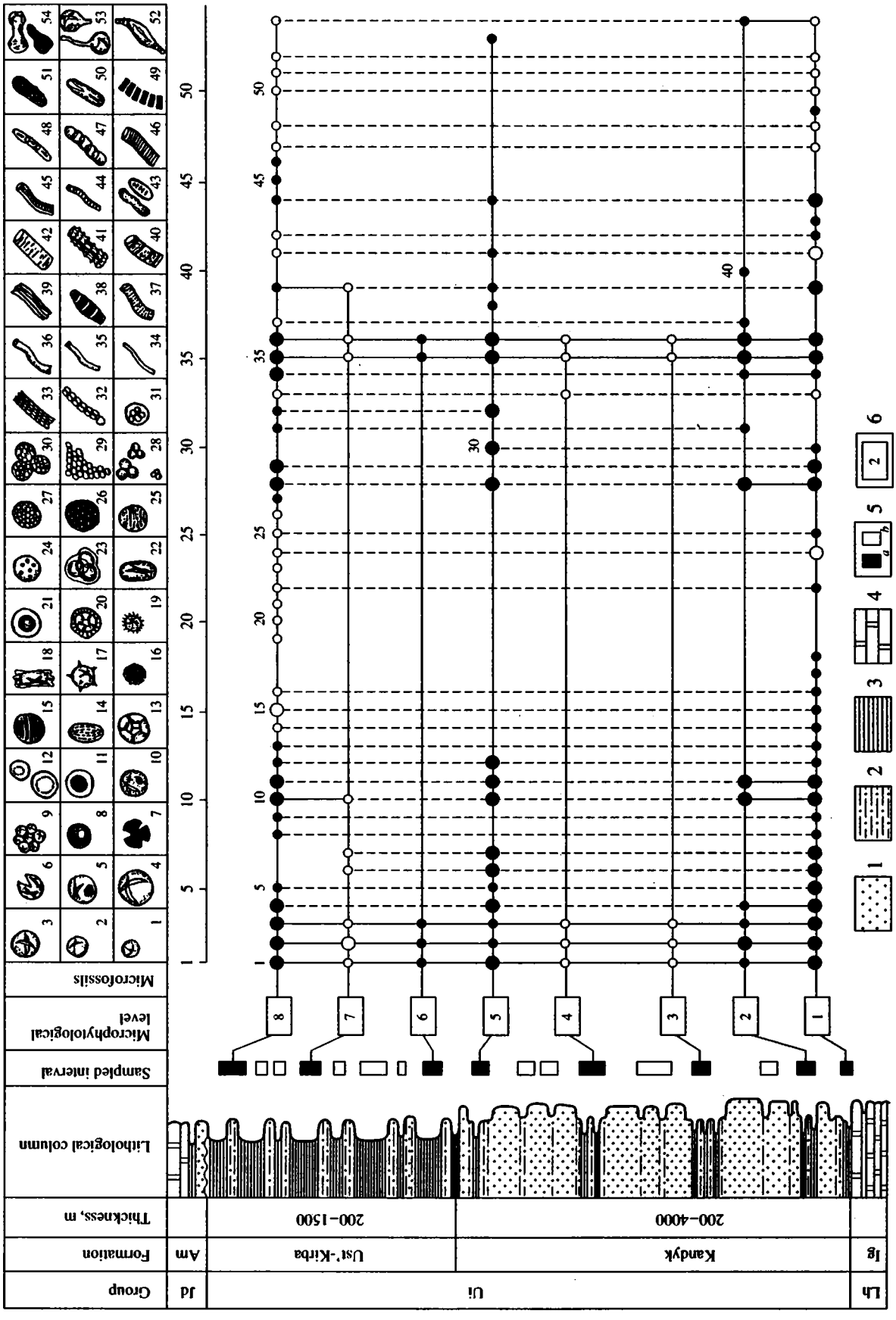
The first level encloses the microfossil assemblage that is most representative of the Kandyk Formation and most diverse among others distinguished in the Ui Group. It includes 34 genera and 62 species of different structure, most of which is known from the underlying Lakhanda deposits. The assemblage is very similar to Lakhanda microbiota in both the taxonomic composition and preservation state (predominance of light-colored thin-walled morphotypes associated with very abundant films of decomposed organic matter). The similarity is so evident that Timofeev (1966, 1969, and others) referred the Kandyk microfossils to the Lakhanda microbiota. According to data of Pyatiletov (1988), which are consistent with our results, forms dominant at the first level are transit taxa widespread in the Riphean, such as *Leiominuscula, Leiosphaeridia, Nucellosphaeridium, Pterospermopsimorpha, Myxococcoides, Coniunctiophycus, Eomycetopsis, Leiotrichoides, Oscillatoriopsis, Archaeoellipsoides* (=*Brevitrichoides*), and some others. The typical Late Riphean *Chuaria* (including very big specimens), *Cucumiforma, Tetrasphaera, Valeria, Simia* (=*Pterospermella*), *Spumosina, Ostiana, Polysphaeroides, Tortunema, Polytrichoides, Navifusa, Digitus, Fabiformis, Arctacellularia, Trachytrichoides, Pellicularia, Aimia, Rugosoopsis*, and *Plicatidium*, are also abundant. The acritarchs of the first level vary in size from 5–6 to more than 1000 μm. Cell diameters of colonial coccoidal forms range from 8–10 to 30–32 μm. Sheaths are from 3–4 to 100 μm wide and wider. Trichomes and trichomelike forms are 6 to 50 μm across.

The Ui assemblages were earlier considered as lacking large Late Riphean acanthomorphic acritarchs abundant in the Lakhanda microbiota (Timofeev *et al.*, 1976; Pyatiletov, 1988; Veis, 1988). Now the first level yielded two varieties of these forms: the large (more than 400 μm) oval *Prolatoforma* with bipolar short cylindrical processes inside filmy margins and the smaller (140 μm) *Trachyhystrichosphaera* with long sharpened processes and broad margins. The other, previously unknown Kandyk morphotypes are tabular *Satka* (55–60 μm across), small unnamed "lophostriated" envelopes (24 μm), surface of which is decorated with regular rows of rounded granules (1–2 μm), and coenobial aggregates of *Leiosphaeridia* (envelope sizes 40 to 50 μm), colonial coccoidal *Coniunctiophycus* (cell 16–18 μm in diameter, individual colony 40–60 μm wide, and their aggregates 180–200 μm across), and very large trichomelike forms of oscillatorian type (up to 50 μm wide).

The second level yields the essentially different microfossil assemblage. Forms dominant here are medium-sized thin-walled *Leiosphaeridia* (100–150 μm, the total size range is 30–290 μm). Larger *Chuaria* (310–550 μm) occur less frequently. These forms associate with the following taxa: relatively abundant *Nucellosphaeridium* and *Pterospermopsimorpha* (28–96 μm in diameter); very large peculiar aggregates of few or many envelopes inside a single sheath ex. gr. *Eogloeocapsa* (internal envelopes 60–70 μm, the entire specimen up to 170–190 μm in size); colonial coccoidal *Myxococcoides* (cell diameters 8–12, 14–16, and 22–24 μm); smooth-walled sheaths of *Archaeotrichion, Eomycetopsis*, and *Leiotrichoides* (1, 3–4, and 12 μm wide; individual and interwoven filaments of the second taxon were found); *Siphonophycus* with indistinct transverse ornamentation (30 μm wide); single tubular *Rectia* (40–46 μm wide); and peculiar complicated forms comparable with the smallest specimens of *Amoeboidium* (maximal size 110 × 170 μm).

The third and fourth levels enclose microfossil assemblages of similar though impoverished taxonomic composition (7 species of 4 genera and 8 species of 5 genera, respectively). Dominating forms are small acritarchs *Leiominuscula* (up to 10 μm) and *Leiosphaeridia* (30–70 μm, less frequently up to 200 μm), which associate with sheaths of *Eomycetopsis* (3–4 μm) and *Leiotrichoides* (8–10 and 18–20 μm). In addition, single relatively narrow (10–20 μm) tubular forms with transverse cell rows, which are similar to *Polysphaeroides*, were found at the fourth level.

The fifth level bears a very representative microfossil assemblage (18 genera and 26 species). Microfossils here are only slightly less abundant and diverse than at the first level. The dominant forms are small- to medium-sized acritarchs, colonial coccoidal microfossils, and sheaths. Acritarchs are represented by


Group
Formation
Thickness, m
Lithological column
Sampled interval
Microphytological level
Microfossils
Lh
lg
Ui
Kandyk
Ust'-Kirba
Am
Jd
200–4000
200–1500
1
2
3
4
5
6
7
8
30
40

Leiosphaeridia (30–250 μm; envelopes of 30–70 μm in diameter are more frequent), *L. bicrura* Jank., *L. ternata* (Tim.), and *L. incrassatula* Jank. included. Associated taxa are *Leiominuscula* (up to 10 μm), *Nucellosphaeridium*, *Pterospermopsimorpha* (up to 100 μm), *Simia* (60 μm), and *Chuaria* (up to 520 μm). The colonial microfossils are represented by very abundant subspherical to ellipsoid colonies of *Myxococcoides* (cell diameters 8–10 and 14–16 μm, colonies 80 × 100 μm in size), by aggregates of presumably allied *Coniunctiophycus* whose cells and colonies are of similar sizes, and by uniserial chains of cells (12 μm across) similar to *Sphaerocongregus*. Filamentous forms are dominated by *Eomycetopsis* (2 or 3 μm) and *Leiotrichoides* (6–12 and 20–23 μm). They occur together with multiserial sheaths *Polytrichoides* (up to 30 μm, individual filaments 4 μm wide), tubular *Rugosoopsis* with "toothed" edge (36 μm wide), and transversely ornamented tapering *Siphonophycus attenuatum* A. Weiss (up to 60 μm wide). The filamentous forms are *Oscillatoriopsis* (6 and 10–14 μm across) and *Palaeolyngbya* (sheaths 16–18 to 30 μm wide, and trichomes 4 to 8 μm wide). The forms of complex structure include *Germinosphaera*, the presumable germinating structures of ancient algae or fungi with a single short process (diameter of envelope 130 μm, process 10 μm long), and smaller *Caudosphaera* with a very long process (envelope 40–60 μm across, process up to 250 μm long). The peculiar, occasionally branching thalluses with rounded thickened terminations (16 and 30–40 μm wide) were also recognized

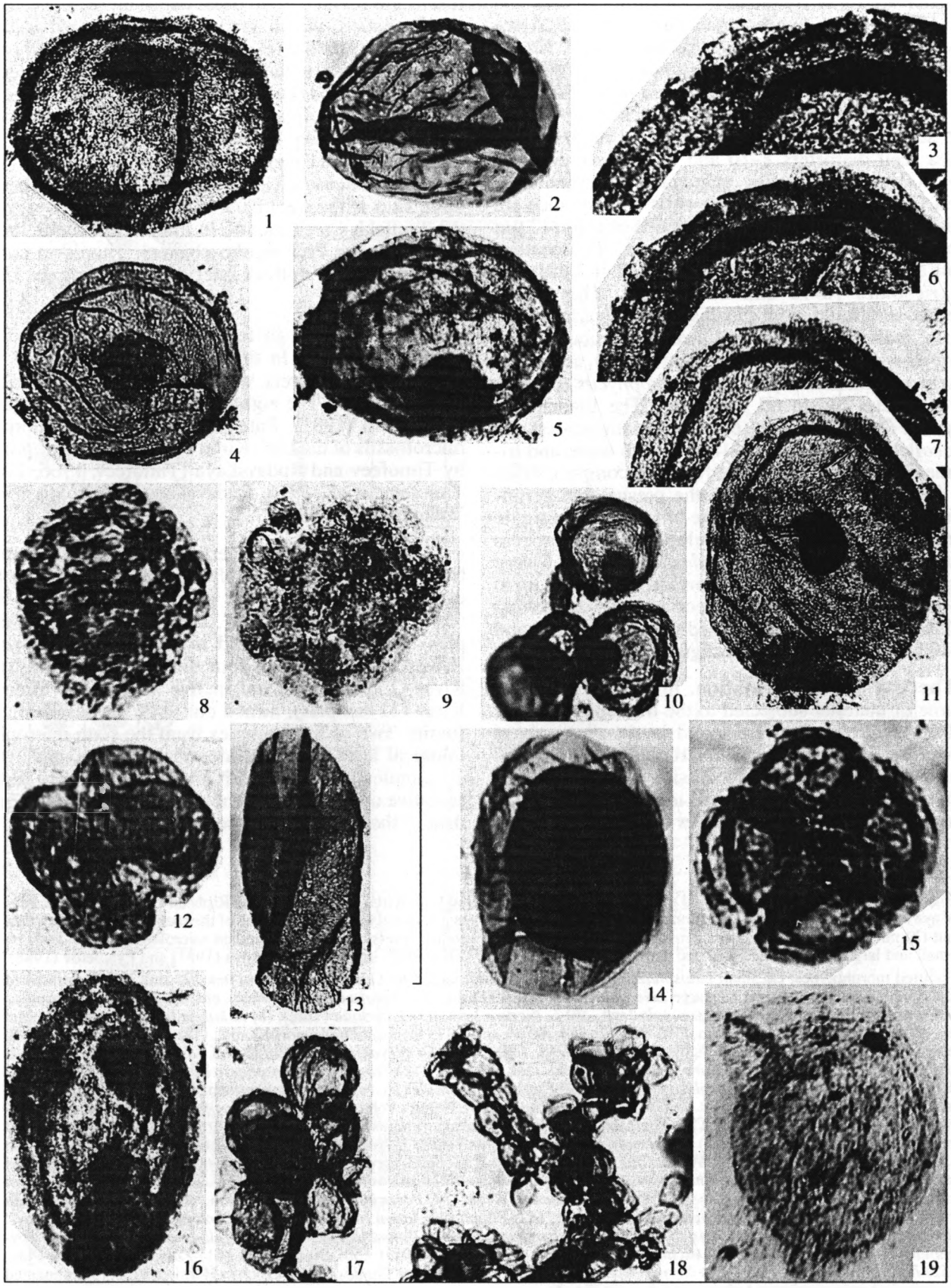
The Ust'-Kirba Formation. In spite of the great volume of material subjected to the microphytological analysis, microfossils were found in six sites only at three arbitrarily established microphytological levels (6th to 8th of the general succession). The sixth level (about 20 m thick) is established in a single outcrop at the right bank of the Yudoma River between the mouths of the Kary-Ytyga and Ulakhan-Ytyga rivers. Microfossils are found here in basal dark-colored deposits of the formation, which grade upward into red-colored ones. The seventh level of imprecise stratigraphic position corresponds to the 20- to 30-m-thick upper interval of the Ust'-Kirba (Dalynda) Formation in two adjacent outcrops at the right bank of the Allakh-Yun' River: near the Chertovo Ulovo locality and upstream the Malaya Sakhara Creek (the Osennii Creek) mouth. The eighth level most rich in microfossils is distinguished in three sites at the right bank of the Yudoma River: near the Tyallakh Creek mouth, in the Nuuchaallakh locality, and at the Prokopenko channel. This level corresponds to the 50-m-thick interval terminating the Ust'-Kirba section.

Microfossils of all levels were earlier studied by Pyatiletov (1988). In this work, we investigated the sixth and eighth levels, which were formerly described by Veis (1988). The eighth level was also investigated by Volkova (1981). Rare and simple sphaeromorphic microfossils of the Ust'-Kirba Formation were reported by Timofeev and Rudavskaya (Timofeev, 1966, 1969; Timofeev *et al.*, 1976; Rudavskaya and Frolov, 1974; Volkova *et al.*, 1980).

The newly studied material consists of 66 samples from 14 outcrops of the Ust'-Kirba Formation in the central and northern parts of the Yudoma–Maya belt (the Yudoma, Allakh-Yun', and Belaya rivers). Samples, which were collected in one site at the Belaya River (10 samples), in four sites at the Allakh-Yun' River (13 samples), and in five sites at the Yudoma River (22 samples), turned out to be devoid of microfossils. Two of four samples from the sixth microphytological level contained identifiable forms, and 10 of 17 samples from the eighth level yielded a very representative microfossil assemblage. Below is summary of data of the previous and present studies.

Fig. 3. Distribution of organic-walled microfossils in the Kandyk and Ust'-Kirba formations of the Riphean Ui Group, the Uchur–Maya region: (1) sandstone, (2) siltstone, (3) mudstone, (4) dolomites; (5) closely sampled intervals of the section with (*a*) and without (*b*) microfossils; (6) number of microphytological level. Occurrence levels of rare and abundant microfossils is indicated by small and larger circles. Data sources: dark circles, this study and Veis (1988); light circles, Volkova (1981) and Pyatiletov (1988). Pictured morphotypes: (1) small *Leiosphaeridia* (less than 70 μm), including *Leiominuscula* (less than 10 μm), (2) medium-sized *Leiosphaeridia* (70–150 μm), (3) large *Leiosphaeridia* (150–300 μm), (4) *Chuaria circularis* Walc., emend. Vidal, Ford (envelope >300 μm), (5) *Leiosphaeridia incrassatula* Jank., (6) *L. bicrura* Jank., (7) *L. ternata* (Tim.), (8) *L. kulgunica* Jank., (9) coenobial aggregates of small *Leiosphaeridia*, (10) *Nucellosphaeridium* spp., (11) *Pterospermopsimorpha* spp., (12) *Simia annulare* (Tim.), emend. Mikh. and/or *S. nerjenica* A. Weiss, (13) *Satka* sp., (14) *Cucumiforma vanavaria* Mikh. emend. Mikh., (15) *Valeria lophostriata* Jank., (16) *Spumosina rubiginosa* (Andr.), emend. Jank. et Medv., (17) *Trachyhystrichosphaera aimica* Herm., (18) *Prolatoforma aculeata* Mikh., (19) *Baltisphaeridium* sp., (20) *Cymatiosphaeroides* sp. (=?*Cymatiosphaera* sp.; Volkova, 1981), (21) *Granomarginatasphaera judomica* Pjat., (=*Granomarginata squamacea* Naum.; Volkova, 1981), (22) *Navifusa majensis* Pjat., (23) *Tetrasphaera antiqua* Tim. et Herm., (24) *Aimia* spp., (25) ?lophostriated envelopes consisting of small granules (1 or 2 μm) arranged in rows, (26) *Tasmanites* sp., (27) *Sphaerophycus parvum* Schopf, (28) *Myxococcoides* spp., (29) *Ostiana microcystis* Herm., (30) *Coniunctiophycus totticus* A. Weiss, (31) *Eogloeocapsa* sp., (32) *Sphaerocongregus variabilis* Moorm., (33) *Polysphaeroides contextus* Herm., (34) *Archaeotrichion contortum* Schopf, (35) *Eomycetopsis robusta* Schopf, emend. Knoll, Golubic, (36) *Leiotrichoides* spp., (37) *Siphonophycus* sp., (38) *S. attenuatum* A. Weiss, (39) *Polytrichoides lineatus* Herm., emend. Herm., (40) *Rectia costata* (Jank.), (41) *Rugosoopsis tenuis* Tim. et Herm., (42) *Plicatidium latum* Jank., (43) *Brevitrichoides* spp. and/or *Archaeoellipsoides* spp., (44) *Oscillatoriopsis* spp., (45) *Palaeolyngbya* spp., (46) *Botuobia* spp., (47) *Arctacellularia* spp., (48) *Trachytrichoides ovalis* Herm., (49) broad trichomelike structures (more than 50 μm), (50) *Digitus* spp., (51) *Fabiformis porosus* Pjat., (52) *Pellicularia tenera* Jank., (53) *Germinosphaera tadasii* A. Weiss and/or *Caudosphaera expansa* Herm. et Tim., (54) unnamed forms of complex structure (other symbols as in Fig. 2).

Plate III



The sixth level is characterized by an impoverished microfossil assemblage (4 genera, 8 species, according to Pyatiletov, 1988) that is represented in our material only by acritarchs *Leiosphaeridia* (envelopes ranging in diameter from 30 to 180, sometimes, to 250 μm) and by *Eomycetopsis* and *Leiotrichoides* sheaths (3–4 and 8–12 μm wide).

The seventh level shows a higher taxonomic diversity of microfossils (7 genera and 13 species, according to Pyatiletov, 1988). The assemblage mainly consists of acritarchs *Leiominuscula* (up to 10 μm) and *Leiosphaeridia* (up to 280 μm in diameter), *L. bicrura* Jank. and *L. ternata* (Tim.) included, and also of *Nucellosphaeridium* (100–150 μm) and sheaths of simple narrow *Eomycetopsis* (3 or 4 μm), wider *Leiotrichoides* (10–12 and 20–22 μm), multiserial *Polytrichoides*, and some problematic forms.

The eighth level. All who studied microfossils of this level (see references above) emphasized their abundance and high taxonomic diversity. Pyatiletov (1988, p. 55) noted most categorically that "...numerous samples yielded almost all microfossils of the Kandyk Formation in association with *Cymatiosphaera* sp., *Tasmanites* sp., *Sphaerophycus parvum* Schopf, and *Siphonophycus kestron* Schopf." In total, he recognized 38 genera and 60 species, among them *Chuaria, Valeria, Tetrasphaera, Cucumiforma, Spumosina, Simia, Navifusa, Leiovalia, Digitus, Fabiformis, Tortunema, Polytrichoides, Polysphaeroides, Arctacellularia, Ostiana, Pellicularia, Rugosoopsis, Plicatidium*, very wide *Oscillatoriopsis* (=*Botuobia*), supposedly chitinoid *Majasphaeridium*, and some other remarkable morphotypes characteristic of the Lakhanda Group and Kandyk Formation.

Volkova (1981), who concentrated her attention on acritarchs and additionally analyzed other morphotypes, was more cautious in her evaluation of the late Ust'-Kirba microfossils. In the Nuuchaallakh locality, she found 14 species of simple acritarchs representing 10 genera. *Cymatiosphaera* sp. and *Granomarginata* sp. (probably, *Cymatiosphaeroides* and *Granomarginatasphaera* in recent classification) occurring among them were previously unknown from the Riphean deposits, like *Valeria, ?Baltisphaeridium* sp., and *?Tasmanites* sp. These forms were found together with closely spaced aggregates of coccoidal cells of medium (10–20 μm) and small (2 or 3 μm) size, and with filamentous nonceptate microfossils 2–15 μm wide.

The newly obtained data are consistent with previous results. However, our samples did not yield the forms characteristic of the Lakhanda–Kandyk interval, such as *Arctacellularia, Rugosoopsis, Pellicularia*, and *Plicatidium* reported by Pyatiletov and *Cymatiosphaera, Baltisphaeridium*, and *Tasmanites* found by Volkova. Our acritarchs vary in the size from 8–10 to 545 μm. Dominant them are transit *Leiominuscula, Leiosphaeridia, Chuaria, Nucellosphaeridium, Pterospermopsimorpha*, and *Simia*, but we also identified *Satka* (140–150 μm) and coenobial aggregates of *Leiosphaeridia* (some envelopes 40 μm in diameter) in this group. Colonial coccoidal forms are represented by small shapeless *Myxococcoides* (cell size is 8–10, 14–16, and 20–22 μm), long thalluses *Ostiana* (cell diameter up to 30 μm), dense subspherical *Sphaerophycus* (cell diameter 3 μm; size of colony 30–40 μm), chainlike *Sphaerocongregus* (2 or 3 μm), and *Eogloeocapsa* bodies enclosed into a single sheath (of internal bodies 16–18 and 30–32 μm in size, and the sheath about 100 × 150 μm in dimensions). The filamentous forms are the structureless sheaths of *Archaeotrichion* (1–1.5 μm), *Eomycetopsis* (2 or 3 μm), *Leiotrichoides* (8–13 μm) (individual and closely interwoven specimens of two former morphotypes were found), and multiserial *Polytrichoides*. Less abundant trichomes belong to *Oscillatoriopsis* (7 and 14 μm wide), filamentous *Palaeolyngbya* (width of sheaths 20 μm, that of trichomes 5 or 6 μm), and rare *Botuobia* (40–50 μm). The accessory components are rare elongate-ellipsoidal envelopes of *Digitus* (8 × 75 μm), which may be varieties of common Riphean akinete-like forms of the genus *Archaeoellipsoides*.

Thus, the studied new representative material from the Ui Group (56 samples with microfossils collected at eight microphytological levels of 11 sites) demon-

Plate III. Microfossils from the Kandyk and Ust'-Kirba formations of the Upper Riphean Ui Group, the Uchur–Maya region.

All specimens pictured here are stored at the Geological Institute of Russian Academy of Sciences, Moscow.

(1–11, 13, 16–19) the Kandyk Formation: (1, 3, 4, 6, 7, 11, 13, 16, 19) first microphytological level, the Bol'shoi Kandyk River mouth area, (2, 5, 8–10, 17, 18) fifth microphytological level, the right bank of the Maya River, 4–5 km downstream the Bol'shoi Kandyk River mouth; (12, 14, 15) the Ust'-Kirba Formation, eighth microphytological level, the Yudoma River, Nuuchaallakh Locality.

Scale bars: 250 μm for (1, 5, 13); 200 μm for (2, 4, 19); 800 μm for (3); 600 μm for (6); 400 μm for (7, 16); 120 μm for (8, 9, 17); 110 μm for (10); 300 μm for (11); 40 μm for (12, 15); 100 μm for (14); and 80 μm for (18).

(1, 4, 5) *Nucellosphaeridium nordium* (Tim.): (1) sp. no. 111/1138, (4) sp. no. 111/1124, (5) sp. no. 7952/2499; (2, 13) *Leiosphaeridia jacutica* (Tim.), emend. Mikh. et Jank.: (2) sp. no. 8952/3112, (13) sp. no. 111/1180; (3, 6, 7, 11) *Prolatoforma aculeata* Mikh.: (3, 6, 7) magnified fragments of ellipsoidal envelope with bipolar short subcylindical processes, sp. no. 111/1156; (8, 9, 12) *Myxococcoides* spp.: (8) sp. no. 7952/2422, (9) sp. no. 7952/3117, (12) sp. no. 8061/1532; (10) *Nucellosphaeridium minutum* Tim., accumulation of envelopes, sp. no. 7952/3117; (14) *Pterospermopsimorpha pileiformis* Tim., emend. Mikh., sp. no. 8061/1544; (15) *Tetrasphaera antiqua* Tim. et Herm., sp. no. 8061/1599; (16) *Cucumiforma vanavaria* Mikh., sp. no. 111/1221; (17) *Leiosphaeridia crassa* (Naum.) emend. Jank., aggregate of several envelopes, sp. no. 8952/3147; (18) *Ostiana microcystis* Herm., sp. no. 8952/3176; (19) *Trachyhystrichosphaera aimica* Herm., sp. no. 111/2293.

Plate IV

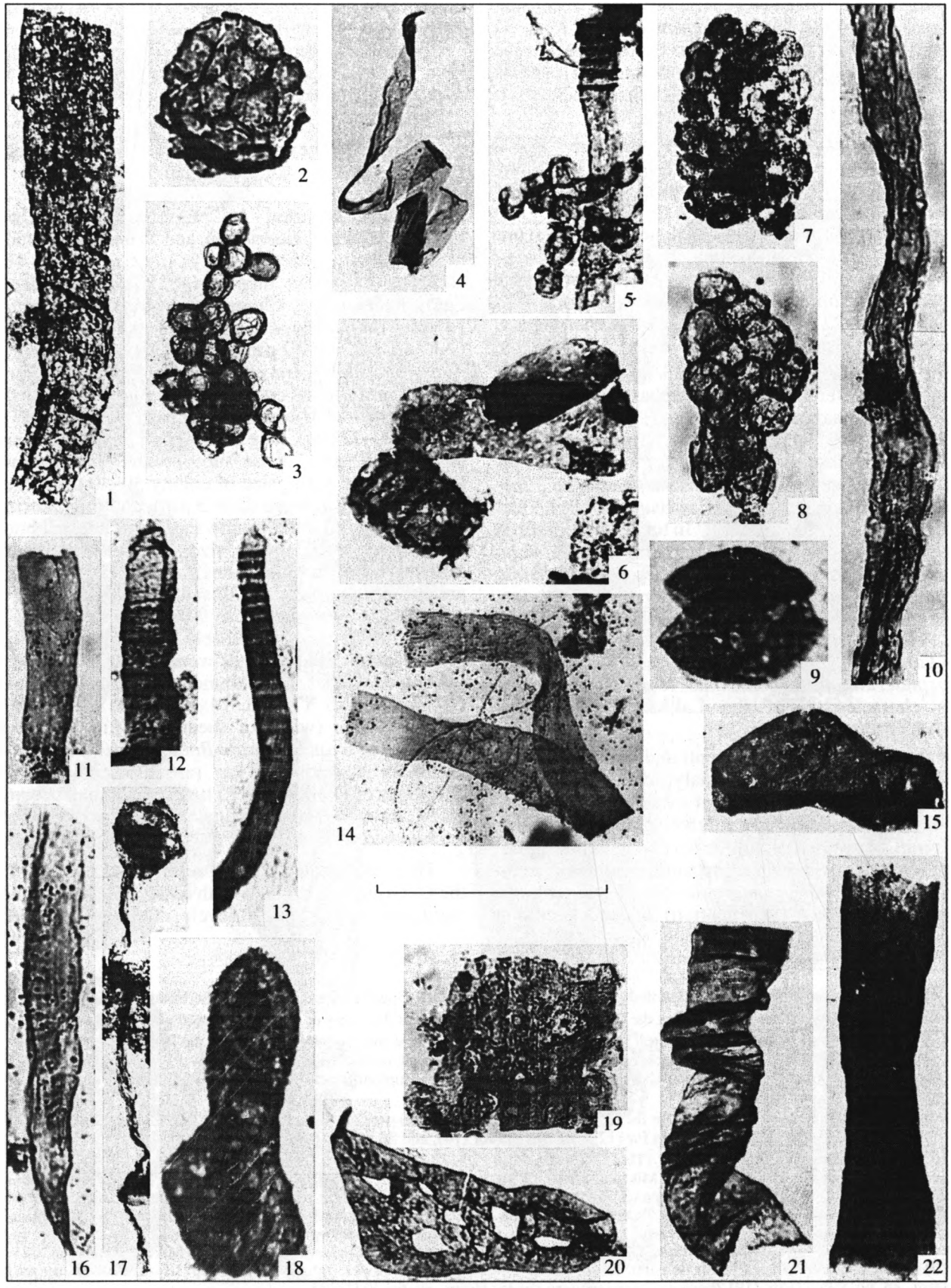


Plate IV. Microfossils from the Kandyk and Ust'-Kirba formations of the Upper Riphean Ui Group, the Uchur–Maya region.

All specimens pictured here are stored at the Geological Institute of Russian Academy of Sciences, Moscow.

(1, 2, 4–11, 14–22) the Kandyk Formation: (1, 9) first microphytological level, the Bol'shoi Kandyk River mouth area, (18) second microphytological level, the Belaya River near the Petrov-Yurege Creek mouth, (2, 4–8, 10, 11, 14–17, 19–22) the microphytological level, the right bank of the Maya River, 4–5 km downstream the Bol'shoi Kandyk mouth; (3, 12, 13) the Ust'-Kirba Formation, eighth microphytological level, the Yudoma River, Nuuchaallakh Locality.

Scale bars: 120 μm for (1, 18); 80 μm for (2, 7, 8, 16); 240 μm for (3, 5); 300 μm for (4); 100 μm for (6, 21, 22); 50 μm for (9, 19); 160 μm for (10, 11, 14, 15, 17, 20); and 60 μm for (12, 13).

(1) a fragment of broad heavily corroded ribbonlike thallus, sp. no. 111/1189; (2, 3, 7, 8) *Myxococcoides* spp.: (2) sp. no. 8952/3166, (3) sp. no. 8064/1552, (7) sp. no. 7952/2216, (8) sp. no. 7952/2224; (4, 14) plastic smooth-walled ribbonlike thalluses: (4) sp. no. 8952/3154, (14) sp. no. 8952/3290; (5) an aggregate of coccoidal cells ex gr. *Myxococcoides* and a fragment of filamentous sheath with one terminal swelling indicating a cell structure of the trichome, sp. no. 8952/3142; (6) *Archaeoellipsoides* spp., sp. no. 8952/3228; (9) *Leiosphaeridia bicrura* Jank., sp. no. 8064/1548; (10) thallus with weakly pronounced longitudinal sculpture ex. gr. *Pseudodendron*, sp. no. 8952/3137; (11) *Asperatofilum experatus* Herm., sp. no. 8952/3186; (12) *Siphonophycus* sp., sp. no. 8064/1608; (13) *Oscillatoriopsis media* Mendelson et Schopf, sp. no. 8064/1122; (15) *Leiosphaeridia jacutica* (Tim.), emend. Mikh. et Jank., deformed envelope, sp. no. 8952/3161; (16) *Palaeolyngbya catenata* Herm., sp. no. 8952/3277; (17) *Caudosphaera expansa* Herm. et Tim., sp. no. 8952/3199; (18) *Amoeboidium asimmetricum* Tim. et Herm., sp. no. 7933/1027; (19) broad trichomelike remnant ex gr. *Botuobia*, sp. no. 7952/3144; (20) unnamed plastic form of unusual structure, sp. no. 8952/3181; (21) broad thallus with weakly pronounced transverse sculpture, sp. no. 8952/3245; (22) dense filamentous ?sheath with thin light zones near broaden terminations, sp. no. 8952/3267.

strated a considerably higher taxonomic diversity of microfossils in question. It should be emphasized that this diversity increase resulted from discovery of microfossils known from lower stratigraphic intervals of the Uchur–Maya and other regions but unrecognized previously in the deposits under consideration. These are not new but transit Riphean (partly Proterozoic) taxa *Eogloeocapsa, Coniunctiophycus, Siphonophycus, Palaeolyngbya, Satka, Germinosphaera*, and specific Late Riphean *Rectia, Caudosphaera*, very wide *Botuobia*, large trichomelike forms, *Amoeboidium, Sphaerocongregus, Siphonophycus attenuatum* A. Weiss, and acanthomorphic acritarchs *Trachyhystrichosphaera* and *Prolatoforma*. With due account for new results, the Kandyk and Ust'-Kirba assemblages of microfossils include about 40 genera and 60 species of different structure, thus being among the most abundant and diverse Proterozoic microbiotas. The confident paleontological characteristics of the Ui Group is especially important for biostratigraphic correlation of this subdivision with other Riphean sections of Siberia and the South Urals in view of new isotope dates for the lower part of the succession (older than 1005 ± 4 and younger than 1025 ± 40 Ma; see Semikhatov *et al.*, 2000).

MAIN INFERENCES

Among two competitive approaches to understanding of the Late Precambrian microbiota evolution, i.e., the geochronological (biochronometrical) and biostratigraphic ones, the former one integrates any microfossil occurrences, including even single specimens, but the direct and indirect information on isotope age of enclosing deposits is of highest priority in this case. Moreover, the succession of microbiotas in the hypo- and stratotypes of the Riphean Eonotheme is considered as having the same significance as the distribution of microfossils in other complete or fragmentary Riphean sections (Knoll, 1994; Knoll and Sergeev, 1995; Sergeev, 1997; Sergeev and Li Seong-Joo, 2001; Yakshin and Faizullin, 1997; and others).

We prefer the alternative approach to chronostratigraphic considerations. It implies that ages of concrete microfossil assemblages, which are important for biostratigraphy, should be evaluated in accordance with the position of these assemblages in the microbiota succession in the Riphean stratotype of the South Urals. This succession can be accepted for the "standard" one by analogy with Phanerozoic successions. Data on facies–ecological peculiarities of host deposits, containing representative regional microbiota successions make the approach more effective and objective (Veis *et al.*, 1998a). It must be emphasized that potentially imperfect microphytological (and any other) characteristics of the Riphean type section may and must be complemented by corresponding data on the Riphean hypostratotype in the Uchur–Maya region, where it was established for this very purpose (Semikhatov and Serebryakov, 1983).

In view of existing differences in correlation of two most important Riphean sections, the procedure under discussion turns into a nontrivial problem that has different solution depending on geochronological or biostratigraphic approach. In this connection, the obtained microphytological characteristics of the Uchur and Ui groups, the initial and terminal members of the Riphean succession in the Uchur–Maya region, are of a great importance. In terms of chronostratigraphy, these data bear not only regional, but global significance as well, because they supplement information on taxonomic diversity and biological connections of chronologically comparable assemblages of microfossils from the Riphean stratotype and objectively elucidate the initial and terminal evolutionary stages of the Riphean microbiota.

A deeper insight into the initial stage can be got through the comparative analysis of the Omakhta

(Uchur) microbiota with that characteristic of the type Lower Riphean Burzyan Group in the South Urals. Organic-walled microfossils occur in all constituent subdivisions of this group, i.e., in the Ai, Satka and Bakal formations (Yankauskas, 1982; *Stratotip...*, 1982; Veis *et al.*, 1990; Mikhailova and Podkovyrov, 1992; and others), whereas the silicified ones are present only in the Satka Formation (Sergeev, 1992).

According to results obtained by many researchers who investigated the materials of different volume and used various classifications (abundant microfossils of 16 genera and 18 species in collection of Yankauskas, scarce specimens of 4 genera and 8 species in the collection of Mikhailova and Podkovyrov, and relatively rare forms of 7 genera and 10 species collected by Veis and his colleagues), organic-walled microfossils of the Burzyan Group represent one and the same microbiota. Principal forms in the type assemblages of microfossils from the Burzyan siliciclastic deposits are small acritarchs *Leiosphaeridia* (usually up to 70 μm, less frequently up to 100 μm), colonial coccoidal *Myxococcoides* (cell diameter up to 26 μm) and filamentous sheaths of *Eomycetopsis* (3–4 μm) and *Leiotrichoides* (up to 10 μm), which associate with les abundant *Nucellosphaeridium, Pterospermopsimorpha, Satka, Eomarginata, Germinosphaera, Palaeoanacystis, Palaeopleurocapsa*, and *Brevitrichoides* (=*Archaeoellipsoides*) (their size does not exceed the ranges indicated above). The organic-walled forms are most abundance in the upper, Bakal Formation of the Burzyan Group.

Silicified microfossils the Burzyan (Satka) assemblage (8 genera, 7 species) are similarly abundant but less diverse than the organic-walled forms (Sergeev, 1992). They include coccoidal *Eogloeocapsa, Gloeodiniopsis* (envelope diameter up to 45 μm), small ellipsoidal *Eosynechococcus* (0.5–3 × 3.5–9.5 μm), filamentous *Gunflintia* and *Eomycetopsis* (up to 7 μm wide), and also larger *Oscillatoriopsis, Palaeolyngbya* (up to 20 μm wide), and *Siphonophycus* (35 μm wide). However, Sergeev (pers. commun.) recently doubted the conclusion that three latter forms occur in deposits of the Burzyan Group.

The comparison of Burzyan and Omakhta microbiotas implies two important inferences. First, these microbiotas show the closest taxonomic similarity characteristic of both the organic-walled and silicified microfossils. As for silicified microfossils known from the Satka Formation, only *Eogloeocapsa, Eosynechococcus*, and *Gunflintia* occurring in the latter have not been identified in the Omakhta deposits (the different preservation degree is taken into account), but this lacuna can be filled, we guess, after the proper investigation of abundant Omakhta cherts. On the contrary, the Omakhta assemblage of organic-walled microfossils is more diverse than the Burzyan one, because *Simia, Konderia, Coniunctiophycus, Palaeopleurocapsa, Polycavita, Stratimorphus, Cyanonema, Filiconstrictosus, Aimia*, and forms of complex morphology are missing from the latter.

Second, the close similarity of the Burzyan and Omakhta microbiotas, in combination with inferable sedimentological parameters, do not allow us to explain their low diversity by the facies–ecological factors, e.g., by the extremely shallow sedimentation settings (Sergeev and Li Seong-Joo, 2001; Semikhatov *et al.*, in press). First of all, the most abundant and diverse microfossils (the thirteenth and fourteenth levels) are enclosed in sediments that accumulated far from the shore of the Omakhta basin, in the subtidal zone with most calm hydrodynamic conditions (Veis and Semikhatov, 1989). Later on in the geological history, this zone was occupied by abundant organisms of large size and complex structure. Under the stable hydrodynamic conditions, gigantic acritarchs and specific coccoidal or filamentous forms could easily penetrate into shallower sedimentation settings than those of the Omakhta and Burzyan basins. This phenomenon was analyzed in detail by consideration of the Late Riphean Derevnya and Miroedikha microbiotas of the Turukhansk Uplift (Petrov and Veis, 1995, Figs. 5 and 6; Veis *et al.*, 1998a, 1998b, Fig. 3) and other Siberian microfossils (Nagovitsin, 2001). Consequently, the absence of large and complicated microfossils or their fragments in the Lower Riphean interval of the Riphean stratotype and hypostratotype is very significant. In addition, the Omakhta microbiota exceeds in abundance and diversity all the Burzyan assemblages taken together. The latter occur in the comprehensively studied Ai, Satka and Bakal sediments, which accumulated from time to time in the moderately deep settings of carbonate platform and siliciclastic shelf below the storm wave base (Maslov, 1997; and others). This fact disagrees with the insufficiently substantiated idea that all known Early Riphean microfossil assemblages of the Uchur–Maya region and South Ural are confined to sediments the shallowest settings. Such facies confinement is proved only for some microfossils of the Bakal Formation (Krupenin, 1999).

Thus, the attempts to interpret the Omakhta and Burzyan microbiotas as impoverished marginal communities, which developed under extreme conditions of shallow-water zones of the Early Riphean basins in contrast to more abundant and diverse coenoses typical of the distal settings of the same basins, seem incorrect from sedimentological and paleoecological points of view. Interpretations of this kind are also invalid in the biostratigraphic context, as it is evident from the comparison of microbiotas under consideration with microfossils from the Ust'-Il'ya and Kotuikan formations of the western Anabar Uplift. Stromatolitic biostratigraphy and differently interpreted isotopic-geochronological and chemostratigraphic data suggest that the latter can be of the Early Riphean (for review and bibliography see Semikhatov, 1995; Gorokhov *et al.*, 2001) or even of the Early Proterozoic age (Pokrovskii and Vinogradov, 1991). Our knowledge on microfossils

from the Anabar succession essentially progressed in the recent time (Yakshin, 1991; Veis and Vorob'eva, 1992; Sergeev, 1993; Veis and Petrov, 1994; Golovenok and Belova, 1984, 1994; Sergeev *et al.*, 1995). Some levels of the Ust'-Il'ya and Kotuikan formations yielded plentiful organic-walled and silicified microfossils, including their very large and complex morphotypes. Biostratigraphic analysis of the latter forms (Veis and Petrov, 1994; Veis *et al.*, 1998a, 2001) implies the following conclusions.

The similar Ust'-Il'ya and Kotuikan microbiotas include the following dominant forms: gigantic acritarchs *Chuaria* (up to 1500 μm across), long thalluses of coccoidal *Ostiana*, chainlike aggregates of *Sphaerocongregus* cells, multiserial sheaths *Polytrichoides*, tubular crinkled-costate *Rectia*, wide trichomes *Botuobia*, branching thalluses *Ulophyton*, and other notable morphotypes. In the Uchur–Maya region, all of them had their first occurrence levels in the Kerpyl' Group (Veis, 1988; Veis and Vorob'eva, 1993). In the South Urals, most indicative of these forms appear above the base of the Upper Riphean Karatau Group and in the local Kuzha Formation provisionally regarded by some workers as a member of this erathem (Veis *et al.*, 2000). In addition, some level of the Lower Kotuikan Subformation yielded some more evolutionary "advanced" morphotypes, which appear beginning from the level of Lakhanda deposits and their correlatives in Middle Siberia, and in the Bederysh Subformation of the Zil'merdak Formation of the Karatau Group in the South Urals (Pyatiletov, 1988; Yankauskas, 1982; *Mikrofossilii...*, 1989; German, 1990). These are the regularly branching *Stiganema*-like multiserial chains of cells surrounded by a sheath, large spiral-annular *Elatera*, gigantic trichomelike forms, branching swollen thalluses of *Archaeoclada*, large articulate microfossils similar to *Eosolena*, filamentous *Vanavarataenia* with attached envelopes, and some other morphotypes. It is noteworthy that in the Anabar sequence the characteristic Lakhanda–Derevnya (Bederysh) morphotypes listed above were found only in clayey interbeds accumulated in local depressions between stromatolite bioherms. They do not occur outside these depressions among the "older" Kerpyl–Bezymyannyi (Nugush) forms, which were detected in the relatively shallow-water lower Kotuikan and deeper water Ust'-Il'ya deposits (Veis *et al.*, 2001).

It is a remarkable fact that different facies of a single stratigraphic level reveal a combination of microorganisms of the Kerpyl and Lakhanda appearance, which were at different evolutionary stages and confined to particular ecological niche. First, this elucidates mechanism of evolutionary renewals in the highly persistent Riphean microbiota. Important in this case was a preliminary development ("preparation") of new, more progressive organisms in particular local environments of the Riphean basins, which did not replace but simply entered the former community (Veis, 1999; Veis *et al.*, 1998a, 1999). Second, it becomes evident that the microphytological boundary inside the Lower Kotuikan Subformation, which is similar to the pre-Lakhanda one, represents a special case. It corresponds to an "inner," but not the "outer" boundary of the corresponding megacycle (Veis *et al.*, 1998a). In the Anabar region, this boundary is marked by renewals in composition of trichomelike and complex morphotypes, and partly by taxonomic changes in assemblages of colonial coccoidal microfossils and sheaths, but the corresponding event remained intact the acritarch group. We suspect that a careful microphytological study of the upper Riphean beds in the Anabar region will result in a firm positioning the pre-Lakhanda level, which is so far recognizable here from the aforementioned indirect indications, at the base of the Yusmastakh megacycle. Accordingly, we suggests that the Ust'-Il'ya–Kotuikan organic-walled microfossils are much younger from the viewpoint of biostratigraphy, than the Uchur–Burzyan microbiota, and are correlative only with microfossils from the Kerpyl, Bezymyannyi, and post-Yurmatinian beds (Veis *et al.*, 2000).

Available data on silicified microfossils from the Kotuikan Formation (Yakshin, 1991; Golovenok and Belova, 1984; Sergeev *et al.*, 1995; and others) do not contradict this conclusion. More than 20 genera and species are identified now among these microfossils, and the existing difference in evaluation of their taxonomic diversity is, to considerable extent, a consequence of variable understanding of their biological connections and of different accepted classification. The most remarkable taxa amid them (Semikhatov *et al.*, in press) are palmelloid colonies of *Eoentophysalis*, large akinete-like *Archaeoellipsoides* forms (the long axis more than 100 μm), and gigantic unnamed oscillatorian trichomes (up to 85 μm wide). As is known, *Eoentophysalis* colonies are characteristic of the entire Proterozoic (Hofmann, 1976; *Earth's...*, 1983; *The Proterozoic...*, 1992). Large *Archaeoellipsoides* forms enveloping the germinating trichomes were identified under this and other names (*Glenobotrydion, Arctacellularia, Lomentunella, Brevitrichoides*) as dominants of the Late Riphean microbiotas of the Bederysh (Yankauskas, 1982), Derevnya and Miroedikha (Petrov and Veis, 1995; Veis and Vorob'eva, 1998) formations, and of the Neoproterozoic microbiota of the Wynniatt Formation of the Shaler Group in Arctic Canada, whereas wide trichomes represent characteristic morphotypes of the Karatau Group (*Mikrofossilii...*, 1989). It is also important that recent morphological counterparts of all these taxa are widespread (Golubic, 1973; Hofmann, 1976; Kondrat'eva, 1975; *Kurs...*, 1981; and others).

From the strict biostratigraphic standpoint, these facts clearly show that only the Bursyan and Uchur microfossils are undeniably of the Early Riphean age, and that only they can elucidate the pre-Riphean evolutionary events in the microcosm. The important side of the microcosm evolution was determined by ability of all Early Proterozoic organic-walled microorganisms

(*Leiosphaeridia, Myxococcoides, Eomycetopsis*, and *Leiotrichoides*) to develop further and to become noticeably more diverse and/or much larger in the Early Riphean. Despite the close similarity of pre-Riphean and Early Riphean microbiotas, the latter includes a series of newly appeared and important morphotypes: abundant thin-walled *Leiosphaeridia* (including envelopes of *L. bicrura* Jank. with median fissure and *L. kulgunica* Jank. with pylome), marginate *Simia*, tabular *Satka*, spherical *Pterospermopsimorpha*, ellipsoidal *Konderia* with internal bodies, and *Germinosphaera* envelopes with processes. Thus, the earliest Riphean microbiotas of the South Urals and the Uchur–Maya region are similar or even identical to the Early Proterozoic ones in the composition of colonial coccoidal and filamentous forms, but they differ from the latter in a higher diversity, morphological peculiarities, and larger size of acritarchs (up to 230–240 μm versus 60–100 μm). In addition, the Riphean microbiotas include some newly appeared forms of a complex structure.

In other words, even if we restrict the taxonomical diversity of the Early Riphean microbiotas by their taxa found in the Riphean stratotype and hypostratotype only, we can clearly observe such changes in their abundance and taxonomic diversity at the base of this erathem (the biostratigraphic boundary), which cannot be explained only by paleoecological and taphonomic factors. These quantitative and qualitative changes affected the earliest Riphean microfossils, which are thought to represent remains of eukaryotic, predominantly planktonic and probably benthic organisms. The most evident changes in abundance are typical of planktonic *Leiosphaeridia* (various algae cysts; Tappan, 1980; and others), medium-sized forms of which dominate in the Lower Riphean and overlying deposits of the regions under consideration. The taxonomical changes are exemplified by appearance of more complex planktonic cystlike xeromorphic *Satka* and, among presumable benthic forms, of algal and/or fungal oogonia *Pterospermopsimorpha* (German, 1990), allied *Konderia* and *Simia*, and also of *Germinosphaera*, the presumable zoospores of xantophyte algae or germinating structures of lower fungi (*Mikrofossilii...*, 1989; Butterfield *et al.*, 1988, 1994).

Let us consider now the microbiota development at the end of the Riphean Eonothem on the basis of the available material. First of all, we should compare microfossils the Ui Group and underlying Lakhanda beds with microbiota from the Karatau type section of the South Urals (Yankauskas, 1982; *Stratotip...*, 1982; Veis *et al.*, 1990, 2000; Mikhailova and Podkovyrov, 1992) and briefly discuss the distribution of microfossils in the Derevnya–Miroedikha interval of the Turukhansk succession (Veis *et al.*, 1998a, 1998b, 1999) and in the Neoproterozoic rocks of Spitsbergen and northern Canada (Butterfield *et al.*, 1994; Butterfield and Rainbird, 1998).

The Kandyk and Ust'-Kirba microbiotas (Ui Group) considered above demonstrate the high taxonomic similarity; the only difference is that the Kandyk microbiota lacks the Ust'-Kirba forms *Baltisphaeridium* and *Cymatiosphaera*. In addition, both microbiotas consist almost entirely of the taxa inherited from the underlying Lakhanda deposits. The most important forms, common for the Lakhanda and Ui groups, are *Chuaria, Arctacellularia, Navifusa, Fabiformis, Polysphaeroides, Polytrichoides, Trachytrichoides, Aimia, Valeria, Pellicularia, Rugosoopsis, Rectia, Plicatidium*, and *Majasphaeridium*. At the same time, the Ui microbiotas, which include one new-comer *Cymatiosphaera* sp., are devoid of some important Lakhanda taxa. These are *Eomicrocoleus, Ulophyton, Majaphyton, Aimophyton, Lakhandinia, Elatera, Eosolena*, some species of acanthomorphic *Trachyhystrichosphaera*, and numerous fungilike forms and their germinating structures ex gr. *Mucorites* and *Eosacchoromyces*. Owing to the uniform composition of the Lakhanda and Ui microfossil assemblages, they were earlier considered as a single assemblage (Veis, 1988; Pyatiletov, 1988; Semikhatov, 1991).

In the Turukhansk Uplift, where post-Derevnya deposits occupy the same or slightly higher stratigraphic position, the development of Riphean microbiota was different. The decreased diversity of microfossils in the Burovaya Formation, a probable consequence of inadequate microphytological studies and facies peculiarities (Petrov and Semikhatov, 1998; Sergeev, 2001), is restored in the overlying Miroedikha Formation, where some important morphological taxa appear in addition (German, 1990; Veis and Vorob'eva, 1998; Veis *et al.*, 1998b, 1999). It is remarkable that forms common for the Derevnya and Miroedikha formations are very similar to the Lakhanda–Ui microfossils listed above. These are all morphotypes dominating in the Derevnya Formation, except for some complicated forms represented by single specimens. At the same time, the Miroedikha microbiota contains the newly appeared extremely lophostriated *Valeria*, acanthomorphic *Cymatiosphaeroides*, gigantic *Pellicularia* with longitudinal fissure, envelopes similar to *Cerebrosphaera*, characteristic filamentous *Palaeolyngbya sphaerocephala* Herm. et Pyl., typical *Polysphaeroides filiformis* Herm., and plentiful forms of peculiar structure unknown outside the Miroedikha deposits. These data support the known idea on significant changes in composition of the Precambrian microfossil assemblages across the base of the Miroedikha Formation corresponding to a distinct biostratigraphic boundary (Veis, 1988; Veis and Vorob'eva, 1998).

This boundary is not known in the Uchur–Maya region, but, as was suggested earlier (Veis, 1988), the Ui and Miroedikha microbiotas can be conventionally correlated to each other on the basis of similar taxonomic composition and analogous position in the regional microphytological successions. In addition to numerous transit forms in common, the correlation can

be reliably proved by occurrence of similar complicated acritarchs (*Cymatiosphaera* in the Ust'-Kirba microbiota and *Cymatiosphaeroides* in the Miroedikha microbiota), which are considered to be specific of the youngest Riphean microbiota (Volkova, 1981; Knoll, 1994; Butterfield *et al.*, 1994; and others). It should be also noted that, according to the recent data of C-isotope chemostratigraphy, the Miroedikha Formation is younger than the Lakhanda Group (Bartley *et al.*, in press).

At the same time, the appearance of new morphotypes in the Miroedikha microbiota, which are unknown in the considered Riphean deposits of the Uchur–Maya region, can be interpreted as a reliable biostratigraphic evidence for the post-Ui age of this microbiota. This interpretation is favored by taxonomical changes, which affected, like across two other microphytological boundaries of the Riphean (the pre-Kerpyl and pre-Lakhanda ones), pro- and eukaryotic organisms. The typical Miroedikha morphotypes of undoubted cyanobacterial nature, which are unknown from older deposits, are represented by large *Lyngbya*-like filamentous sheaths *Palaeolyngbya sphaerocephala* Herm. et Pyl. with transverse ornamentation, filamentous *Arthrosiphon* with some annulate thickenings, and some characteristic trichomelike structures. Similarly important are cylindrical sheaths *Polysphaeroides filiformis* Herm., which envelope transversely arranged row of cells and can be attributed either to *Stiganema*-like cyanobacteria by some characters, or to the ancient bangiophytes by others.

A considerably greater amount of the Miroedikha forms are most likely eukaryotes. In the phytoplankton group, these are acanthomorphic acritarchs *Cymatiosphaeroides*, peculiar *Cerebrosphaera* that has cerebriform structure of envelope and is similar to *Chuaria* in diagnostic features (Butterfield *et al.*, 1994), and odd, abnormally lophostriated *Valeria*. Benthic eukaryotes are represented by large articulate *Cephalonyx* resembling some protistan (Aseeva *et al.*, 1987) and the Miroedikha morphotypes of complex structure (Veis and Vorob'eva, 1998; Veis *et al.*, 1998b). Some of them shows a great similarity to multicellular algae of different systematic position, which were found in the stratigraphically similar Wynniatt Formation of the Neoproterozoic Shaler Group in the Arctic Canada (Butterfield and Rainbird, 1998).

Above the Zil'merdak deposits of the South Uralian Upper Riphean (Karatavian) type section, organic-walled microfossils were found in all subdivisions, except for the Katav Formation. Mineralized microfossils are detected in the Min'yar Formation only. The microphytological characteristics of the Karatau succession are very irregular (Keller and Yankauskas, 1980; Yankauskas, 1982; *Stratotip...*, 1982; Veis *et al.*, 1990; Mikhailova and Podkovyrov, 1992; Sergeev, 1992). The Riphean assemblages of organic-walled microfossils are most representative at basal levels of transgressive deposits, being much less abundant and diverse in overlying regressive deposits of the Inzer–Min'yar and Uk–Krivaya Luka sedimentary megacycles (Petrov and Veis, 1995; Veis *et al.*, 1999).

Accordingly, the most remarkable Inzer microfossils occurs at the base of the Lower Sim Subformation or Podinzer beds (Shishenyak microbiota), and those of the Uk Formation are concentrated in its thin basal member. Distinctive forms of the Shishenyak microbiota are some characteristic acritarchs and complicated morphotypes unknown below the Sim Formation in the South Urals. These acritarchs are represented by gigantic *Chuaria circularis* Walc. emend. Vidal et. Ford, perforated *Tasmanites ripheicus* Jank., spongy *Spumosina rubiginosa* (Andr.), marginate *Simia simica* (Jank.), and by all main varieties of acanthomorphic *Trachyhystrichosphaera* and *Prolatoforma aculeata* Mikh. Complex morphotypes are fungilike *Eosacchoromyces* and *Nelcanica* of "zonal" structure.

Among the taxa listed above, large porous envelopes of *Tasmanites* and *Spumosina* probably devoid of internal cavity are most remarkable. Their biological affinity are problematic. Riphean *Tasmanites* are usually referred to representatives of Halosphaerales, but some of them are considered as envelopes of the *Leiosphaeridia–Chuaria* type corroded by mineralization. More peculiar *Spumosina* often are regarded as the most typical late Karatavian–Vendian taxa, but occasionally are attributed to diagenetic artifacts (*Mikrofossilii...*, 1989; Hofmann and Jackson, 1994; Burzin, 1996; and others). The appearance of the characteristic Shishenyak forms indicates a certain taxonomical reorganization in the Karatau microbiotas and marks, as we believe, the regional microphytological boundary at the base of the Inzer Formation. The overlying Inzer–Min'yar deposits yield only the most common taxa of the Riphean. The organic-walled forms representing the latter are small- to medium-sized *Leiosphaeridia, Myxococcoides, Eomycetopsis*, and *Leiotrichoides*. Concurrent silicified forms are *Siphonophycus, Eogloeocapsa, Eosynechococcus, Gloeodiniopsis, Oscillatoriopsis*, and others (Sergeev, 1992).

The basal part of the Uk Formation yields a number of transit Late Riphean taxa: filamentous *Polytrichoides, Arctacellularia, Rectia, Pellicularia*, and *Tortunema*, which associate with simple *Leiosphaeridia*. Some of them transit into the Tolparovo Formation, a facies analogue of the Krivaya Luka Formation (Keller *et al.*, 1984; Mikhailova and Podkovyrov, 1992). At this level, they coexist with *Satka, Trachyhystrichosphaera, Valeria, Pterospermopsimorpha*, and *Spumosina*. Thus, the abundance and diversity of the South Uralian Uk microbiota that consists of taxa first appearing at the earlier Muldakaevo and Shishenyak levels of the Riphean stratotype were restored in the post-Inzer–Min'yar time most likely in response to paleoecological changes.

It is impossible now to use microphytological data to correlate unambiguously the Lakhanda–Ui and Burovaya–Miroedikha intervals of the Riphean in Central Siberia with the post-Zil'merdak subdivisions of the Karatau type section in South Urals, because distribution patterns of the most characteristic microfossils are different in two regions. We should remind that the terminal part of the Riphean succession in the Uchur–Maya region yields very similar Lakhanda and Ui microbiotas, and the latter, lacking some complex morphotypes occurring in the Lakhanda beds, shows only insignificant signs of renewal. The much more significant reduction of taxonomic diversity of the Inzer (Sim) microbiota is recorded in the Uk Formation of the South Uralian stratotype of the Riphean. Among dozens of stratigraphically important Shishenyak taxa, including their representative assemblage of acanthomorphic acritarchs, only the transit Late Riphean *Polysphaeroides contextus* Herm., *Tortunema, Polytrichoides, Arctacellularia* (two species), *Recta*, and *Pellicularia* are inherited by the Uk microbiota, and single *Trachyhystrichosphaera parva* Mikh., *Valeria*, and *Spumosina* occur in the Tolparovo (Krivaya Luka) Formation.

The opposite situation is characteristic of the Turukhansk region, where the most abundant and diverse Miroedikha microbiota, the last one in the regional Riphean succession, is separated from older microbiotas by the distinct regional microphytological boundary. Significance of this boundary is emphasized by appearance of *Cymatiosphaeroides, Miroedikhia*, lophostriated *Valeria, Polysphaeroides filiformis* Herm., *Palaeolyngbya sphaerocephala* Herm. et Pyl., *Cephalonyx, Arthrosiphon*, large *Pellicularia* with a longitudinal fissure, and some unidentified morphotypes of unusual structure, which are characteristic of the Miroedikha microbiota. In addition to these new taxa, the latter includes all important taxa (except for *Baltisphaeridium*) of the Ui and less representative Uk microbiotas. At the same time, the important Miroedikha morphotypes listed above are unknown in the Uchur–Maya region and South Urals.

For this reason, the Miroedikha microbiota can be correlated with the representative Shishenyak microbiota of the South Urals, rather than with the impoverished Uk assemblage. The Uk and Tolparovo microfossil assemblages include only the aforementioned transit morphotypes of the Miroedikha microbiota, but comparing the latter with the Shishenyak microbiota, one can see that these transit morphotypes coexist in both with *Chuaria, Pellicularia, Simia, Valeria, Prolatoforma*, and all main varieties of acanthomorphic *Trachyhystrichosphaera*. It is possible, however, that additional microphytological investigation of the Uk–Krivaya Luka interval in the South Urals and Uk Formation in the Uchur–Maya region may essentially correct the above inference.

Such a perspective seems to be promising, because the "unnatural" sudden impoverishment of the post-Sim microbiotas in the Karatau type section is evident from a valuable information on the Late Neoproterozoic microfossils of northeastern Spitsbergen and Nordaustlandet (Knoll *et al.*, 1989; Butterfield *et al.*, 1988; Knoll, 1992, 1994, 1996). These very diverse microfossils are remains of pro- and eukaryotic uni- to multicellular organisms of ancient planktonic and benthic communities.

They were found at many stratigraphic levels of the Hekla Huk and Murchissonfjorden regional successions, which yield predominantly the mineralized microfossils (the Kapp Lord, Hunnberg, Ryssö, Draken, Backlundtoppen, and other formations). The youngest organic-walled forms of the Riphean age are found here in the Svanbergfjellet Formation (Butterfield *et al.*, 1994). In spite of and mainly owing to the surrounding "mineralized" microfossils, these organic-walled forms can easily find their place in the general biostratigraphic succession of the Riphean microbiotas. They can fill the gap in the terminal part of the Riphean succession in Siberia (post-Miroedikha interval) and, to some extent, in the South Urals.

This evident possibility to improve (more exactly, to complete) the microphytological succession became complicated after the Butterfield's cardinal review of classification of the Precambrian microfossils. He revised systematics of many universally acknowledged taxa and/or subsumed them under synonyms, obeying the letter of the law but contradicting the existing traditions. In accordance with the modified classification, the Svanbergfjellet microbiota consists of 63 morphotypes attributed to 56 genera and species (Butterfield *et al.*, 1994). The most characteristic representatives of this microbiota, which are either absent or occur as single poorly preserved (hardly identifiable) specimens in other localities, are colonial coccoidal *Palaeoastrum* and branching filamentous thalluses *Proterocladus* attributed to remains of benthic green algae of Chlorococcales and Siphonocladales, respectively. Biological affinities of other important morphotypes of this microbiota are less certain. First of all, these are undoubted eukaryotes *Valkyria* with complicatedly dissected tissue of internal body, enigmatic *Pseudotawuia*, supposedly cyanobacteria *Pseudodendron* of Scytonemaceae, and phytoplanktonic acritarchs *Cerebrosphaera, Comasphaeridium, Goniosphaeridium, Gorgonisphaeridium, Dictiotidium, Osculosphaera*, and others.

In the Butterfield's opinion, the very specific Svanbergfjellet microbiota that includes the earliest true remains of green algae can represent a discrete evolutionary link (Lagerstätten Chlorophytes) of the Precambrian microcosm. We agree with this idea and propose to establish an important microphytological boundary at the base of this formation containing many new morphotypes. Butterfield is likely of the same opinion, because he considers the Svanbergfjellet microbiota as

the third principal "evolutionary" stage of the general Neoproterozoic microphytological succession (Butterfield and Rainbird, 1998). The earlier first and second stages are exemplified by the Lakhanda microbiota (Lagerstätten Stramenopiles) and by microfossils from the Wynniatt Formation of the Shaler Group in Arctic Canada (Lagerstätten Alveolates), which have been provisionally described some time ago and show close similarity to the Miroedikha microbiota. These three successive evolutionary links of the Late Riphean microbiota, which are distinguishable on the microphytological basis, can be supplemented with the earliest link corresponding to the Kerpyl' microbiota (possible name Lagerstätten Chuariales), and the whole succession seems to be most justified.

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Stratigraphy of Upper Devonian and Lower Carboniferous Deposits in the Angara River Region

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Abstract—Devonian and Lower Carboniferous deposits of the southern part of the Siberian platform (the Angara River region) are considered in this paper. Lithologic description and paleontological characteristics of all Devonian and Lower Carboniferous stratigraphic units and their analogs are presented, and the units are correlated. Geology and tectonics of the region are analyzed in connection with the formation history of studied deposits.

Key words: stratigraphy, Devonian, Lower Carboniferous, correlation, lithology.

INTRODUCTION

Deposits under discussion control prospecting of diamond placers and their sources. It was the time of their accumulation, when most commercial diamond-bearing kimberlite pipes intruded the southeastern part of the Siberian platform (Brokhfogel' *et al.*, 1997). Tuffaceous interbeds, diamond crystals and associated index minerals, found in Lower Carboniferous deposits of the Angara River region, suggest that primary diamond sources can be discovered in the southern part of the Siberian platform as well. Their prospecting here is hampered by different reasons, first of all, by absence of stratigraphic correlation schemes for deposits studied in this work.

The correlation of Lower Carboniferous deposits in the Angara River region is an intricate problem. This is a consequence of their accumulation in three quite different structural-sedimentation (facies) zones (SSZ), such as the Kansk–Taseeva (key sections 1, 12, 20, and 21), Poima–Biryusa (key sections 2–11, 13–15, 23–25), and Angara–Tunguska (key sections 16–19) zones (Fig. 1).

General and thematic geological works, carried out in the past 25 years in different areas of the Angara River region, revealed a series of formations and sequences similar in age to the known ones, but having quite different structure and composition. The main problem we faced at in this study was their stratigraphic correlation.

MATERIALS

Documentation of many mine openings and boreholes in the Kansk and Taseevo regions, which was carried out together with geologists of the Ivanovsk Expedition (Krasnoyarskgeologiya), and data, which were obtained in the course of research work under themes "The Correlation of Carboniferous Deposits of the Tungusska Basin" and "The Study of the Stratigraphy, Lithology, and Paleogeography of Middle–Upper Paleozoic Deposits of the Rybinskii–Vilyui Activation Zone," are used as a basis for this study. The additional important material was obtained during the geological prospecting in the Chuna and Taishet regions, when we collaborated with the Diamond Searching Team (Irkutskgeologiya).

Paleontological and palynological materials presented in reports of "Krasnoyarskgeologiya" and "Irkutskgeologiya" have been vital for correlation of stratigraphic units.

STRATIGRAPHY AND CORRELATION

The formerly accepted stratigraphic schemes of Lower Carboniferous deposits in the Angara River region are now out of date. Important in this respect are new litho- and biostratigraphic generalizations of the Interdepartmental Stratigraphic Committee (ISC) (*Resheniya...*, 1982). Unfortunately, the Angara River region is inadequately represented in new stratigraphic schemes (the basin of the Angara middle course and the Kan–Taseeva depression have been considered only). In fact, the Lower Carboniferous deposits are much more widespread in the Angara River region.

The situation with Upper Devonian deposits, which are exposed only in the Poima–Biryusa SSZ, is much simpler (Pavlov *et al.*, 1990). The Late Devonian time terminated in the Angara River region with the accumulation of a calcareous–terrigenous sequence attributed to the *Kungus Formation* (after the Kungus River, a tributary of Kan River) by analogy with deposits of a similar type and age in the Rybinskii depression. V.F. Dontsov was first to distinguish the Kungus For-

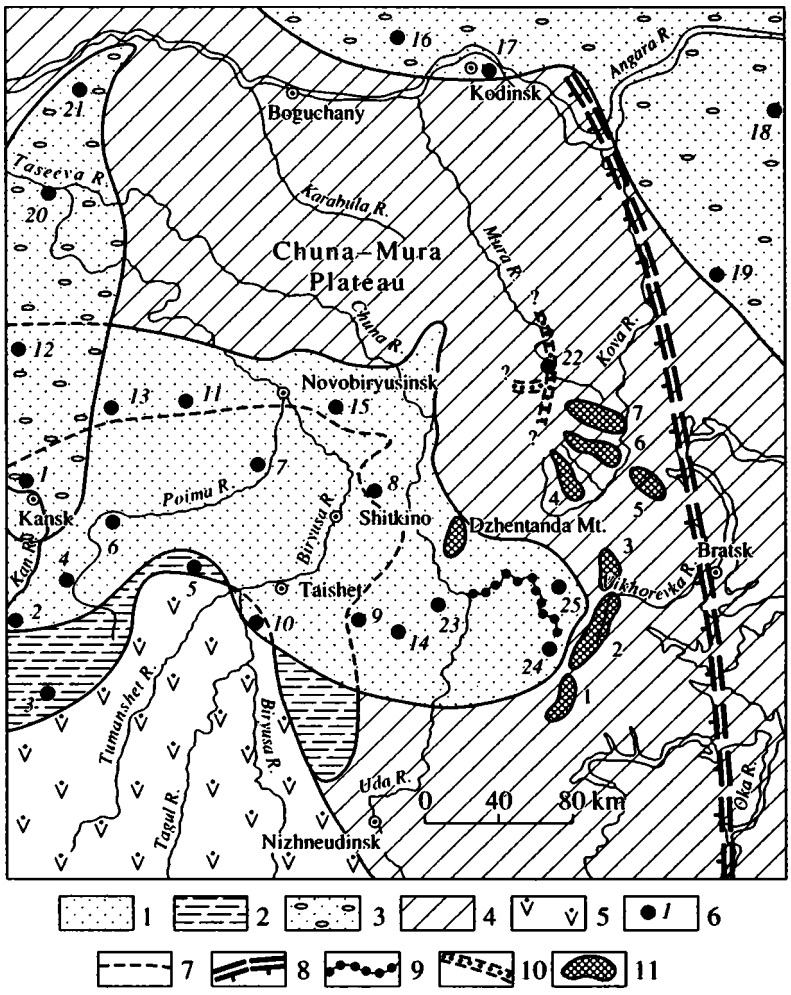


Fig. 1. Sketch map showing localities of key sections of Upper Devonian and Lower Carboniferous deposits in the Angara River region: (1) Tournaisian and Visean, (2) Frasnian and Famennian, (3) Visean and Serpukhovian, (4) Lower Paleozoic, (5) Precambrian and Lower Paleozoic rocks; (6) key section and its number; (7) inferred boundary of Upper Devonian deposits; (8) the Oka–Vikhorevka flexure; (9) the present-day diamondiferous placer at the Chuksha River; (10) the old river bed (deposits of the Mura Formation); (11) present-day ranges composed of Ordovician and Silurian rocks: 1. Seryi Kamen', 2. Konstantinovskii, 3. Katyrminskii, 4. Osinovyi, 5. Mezhzhelonsko–Revunskii, 6. Bumbeiskii, 7. Tyrminskii.

mation in that depression in 1936; the formation name was suggested by A.V. Lesgaft in 1949 (see in Yanov, 1955). In the Angara River region, the formation is mainly represented by red marl with sandstone and siltstone interlayers. Its distinctive feature is occurrence of peculiar beds of marly conglomerate and limestone (calclithite), which are known among local geologists under the name of "roe" marl. Calclithite beds are well traceable from one geologic section to another and represent an important reference horizon.

Deposits of the Kungus Formation overlie transgressively, with a stratigraphic unconformity, the Cambrian, Ordovician, and Middle Devonian deposits (Fig. 2). Conglomerates, from 3 to 30 m thick, are typical almost everywhere of the formation base. The composition of basal conglomerates (pebble of granite, quartz, quartzite, diabase, and schist) varies from the west eastward, where a larger proportion of quartz, quartzite, and chert pebbles was observed. The formation thickness decreases sharply from 264 to 100 m in the same direction (Fig. 2) with parallel reduction of mesocycle numbers from three (key sections 1–5) to two (key sections 6 and 7). Mesocycles well traceable from one section to another facilitate a clear correlation.

The age of the Kungus Formation of the Angara River region is established on the basis of found fauna and flora imprints, and also on results of isotope dating of limestones. During the geological survey of 1959, V.N. Moskalev, L.P. Lyashenko, and V.V. Dedova discovered imprints of placoderms *Dipterus verkeulus* Pand., *Dipterus* sp., *Bothriolepis* sp., *Osteolepis macrolepidotus*, and other representatives of the Osteolepididae family. In addition, the floral assemblage of

Dicranophyton sp., *Hostimella hostimensis* Pet., and *Psilophyton princeps* Daws were also found in the Kungus deposits. K.M. Vainer and N.P. Grigor'ev encountered imprints of fishes of the *Bothriolepis* genus.

Data on the age of limestones in the Angara River region are scarce and controversial. In the vicinity of Kansk, they were previously referred to the Middle Cambrian (Otten, 1929, Yarzhemskaya, 1939, Nechaev, 1949). Doroshenko (1940) correlated them with limestones, which he dated at the Early Silurian. Spores *Archaezonotriletes polimorphus* Naum., which were found later on in carbonates, allowed their assignment to the Devonian (Zharkov, 1960).

The direct isotope dating of sedimentary rocks meets problems as well. The K–Ar age determination is impossible, because authigenic minerals (glauconite and others) have not been found in the studied deposits. Siliceous concretions turned out to be inappropriate for Rb–Sr dating because of a low Rb (0.0003%) and high Sr (0.0024%) contents. Accordingly, the age of bulk limestone samples was determined using Pb–Pb and U–Pb methods (Akulov *et al.*, 1990). These methods are described in works by Moorbath *et al.* (1990) and Bor-ming Jahn (1988). According to Pb–Pb and U–Pb dates obtained, the limestones are 378 to 366 Ma old. The analytical error is up 0.2–0.4%.

Floral and faunal remains mentioned above, and the radiometric dates suggest that deposits of the Kungus Formation accumulated in the Late Devonian (Frasnian to Famennian).

Deposits of the *Charginskaya Formation* that was distinguished by Lesgaft (1948) near the synonymous village, where the type section of the formation occurs at the base of the Lower Carboniferous sequence of the Angara River region (Graizer and Borovskaya, 1964). The formation conformably overlies the Kungus beds. Whitish gray sandstone, locally grading into red siltstone or coarse conglomerate with pebbles of sedimentary and metamorphic rocks mark the formation base.

Two mesocycles, which are well traced in all sections of the Angara River region, are distinguished in the formation structure. In some areas, the first mesocycle predominantly (60%) consists of limestone that is massive, characterized by pelitomorphic to fine-grained structure. The pelitomorphic variety with calcite and quartz nests includes up to 15% of terrigenous admixtures (angular fragments of quartz and feldspar up to 0.3 mm size in size). Streaks of well crystallized calcite are encountered. In fine-grained limestone, terrigenous material is represented by quartzite, chert, feldspar, and muscovite grains varying in size from 0.05 to 0.75 mm and constituting up to 30% of the total rock volume. Grains are acute-angled to subangular.

Sandstones from the lower part of the formation have different cement composition. There are white, fine-grained (sugary), poorly cemented sandstones with argillaceous–carbonate cement and yellow-gray varieties with amorphous siliceous cement. The both varieties are of oligomictic quartz, transitional to monomineral type. They are composed predominantly of quartz (up to 85%) and include about 10% of feldspar and up to 5% of grains of siliceous, clayey, and volcanic rocks. The basal or interstitial cement represents up to 40% of the rock volume. Grains are well sorted. The heavy fraction of accessory minerals is of a constant composition and predominantly consists of garnet (80–89%) and staurolite (2–8%). Pink to lilac–pink pyrope is dominant among garnets of the almandine series. The second mesocycle is composed of irregularly alternating limestone, calcareous siltstone, and mudstone beds, constituting 2- to10-m-thick members. White limestone beds with abundant rounded concretions of reddish-brown cherts are peculiar of the middle part of the sequence. Red calcareous sandstone beds in the second mesocycle consist of quartz (87–90%), feldspar (3–9%), and grains of clayey and effusive rocks (5–6%). Their calcite cement is of basal, or rarely of interstitial and contact types. Minerals of the heavy fraction are garnet (65–70%), ilmenite (15–20%), leucoxene (4–5%), zircon (5–9%), rutile (3–4%), tourmaline (2–3%), and staurolite (1.0–1.5%). The formation thickness varies from 60 to 180 m.

The age of the Charginskaya Formation in the Angara River region was determined by Lebedev (1963) on the basis of fossil flora found near the Village of Voznesenki. In his collection, A.P. Anan'ev identified species *Asterocalamites scrobiculatus* Zeitl. and *Heleniella theodorii* Zal. As he inferred, the formation with such plant remains is an analog of the Tournaisian Bystraya Formation of the Northern Minusinsk depression. In addition, Parfenov (1961) found fish-scale imprints of *Strepsodus siberiacus* Chab. at the base of the formation, which, according to opinion of D.V. Obruchev and A.A. Sergienko, also confirm the Early Carboniferous age of the unit.

Stratigraphic analogs of the Charginskaya Formation are the Chuksha, Saranchet, Baeronovka, and Toporok formations (table).

The Chuksha Formation (after the Chuksha River, a tributary of the Uda River) was first distinguished by V.A. Kozulin *et al.* in 1989 as the synonymous sequence. Rocks of the formation rest with an angular unconformity on Silurian deposits. The rocks make up flat watersheds in the basins of Taz, Chuksha, and Baeronovka rivers. The formation is subdivided into two mesocycles composed of sandstone, siltstone, mudstone, and thin (up to 1 m) conglomerate beds. The total thickness of the formation is 97 m.

Sandstone beds of feldspathic–quartz composition include the light fraction (more than 5%), consisting of siltstone, sandstone, mudstone, and chert grains. An argillaceous cement of sandstone makes up 30–40% of the rock. Imprints of the explicitly redeposited fish fauna, found in sandstone, are attributed to the Silurian by V.I. Byalyi. Feldspathic–quartz, quartz–micaceous,

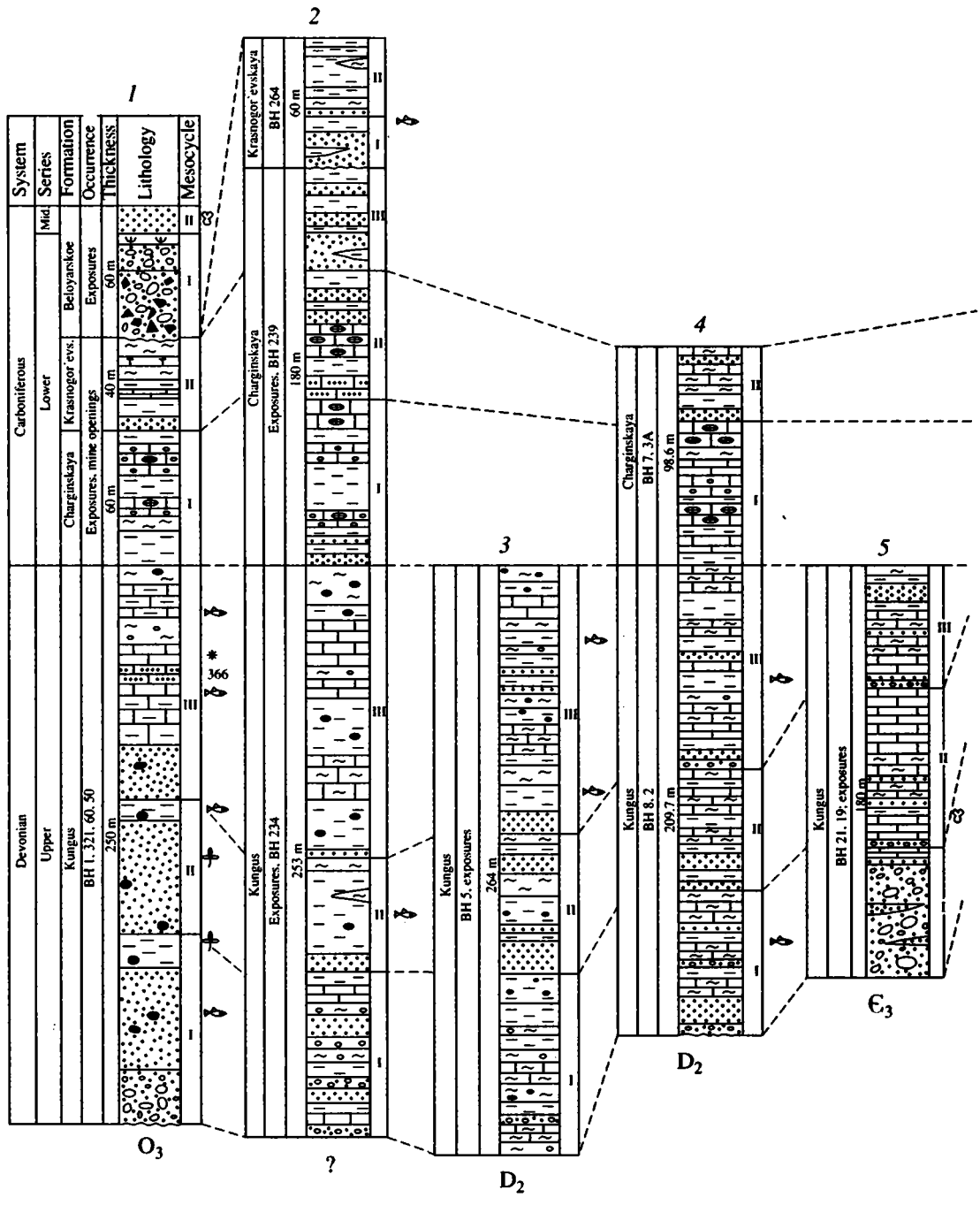


Fig. 2. The correlation scheme for Upper Devonian and Lower Carboniferous stratigraphic units of the Angara River region: (1) sandstone; (2) sandstone with carbonate concretions, (3) sandstone with rare pebbles of blue–green mudstone, (4) sandstone with horizontal bedding, (5) sandstone with cross bedding, (6) sandstone with grus and chippings of siliceous rocks; (7) siltstone; (8) siltstone with carbonate concretions; (9) calclithite with siltstone pebbles; (10) siltstone with individual pebbles of blue–green mudstone; (11) mudstone; (12) calclithite with limestone pebble; (13) siltstone and mudstone lenses; (14) conglomerate; (15) conglomeratic breccia with poorly rounded chert and chalcedony fragments; (16) redeposited products of the kaolin weathering crust; (17) tripoli; (18) tuff; (19) tuffaceous siltstone; (20) tuffaceous mudstone; (21) tuffaceous siltstone with limestone pebbles; (22) limestone; (23) calcareous siltstone; (24) calcareous sandstone; (25) limestone with cherty (chalcedony) concretions; (26) silicified limestone; (27) marl; (28) limestone with stromatolitic structures; (29) flora, (30) fauna, (31) fish remains, (32) spore and pollen; (33) radiological age (*1–25*, key sections by numbers).

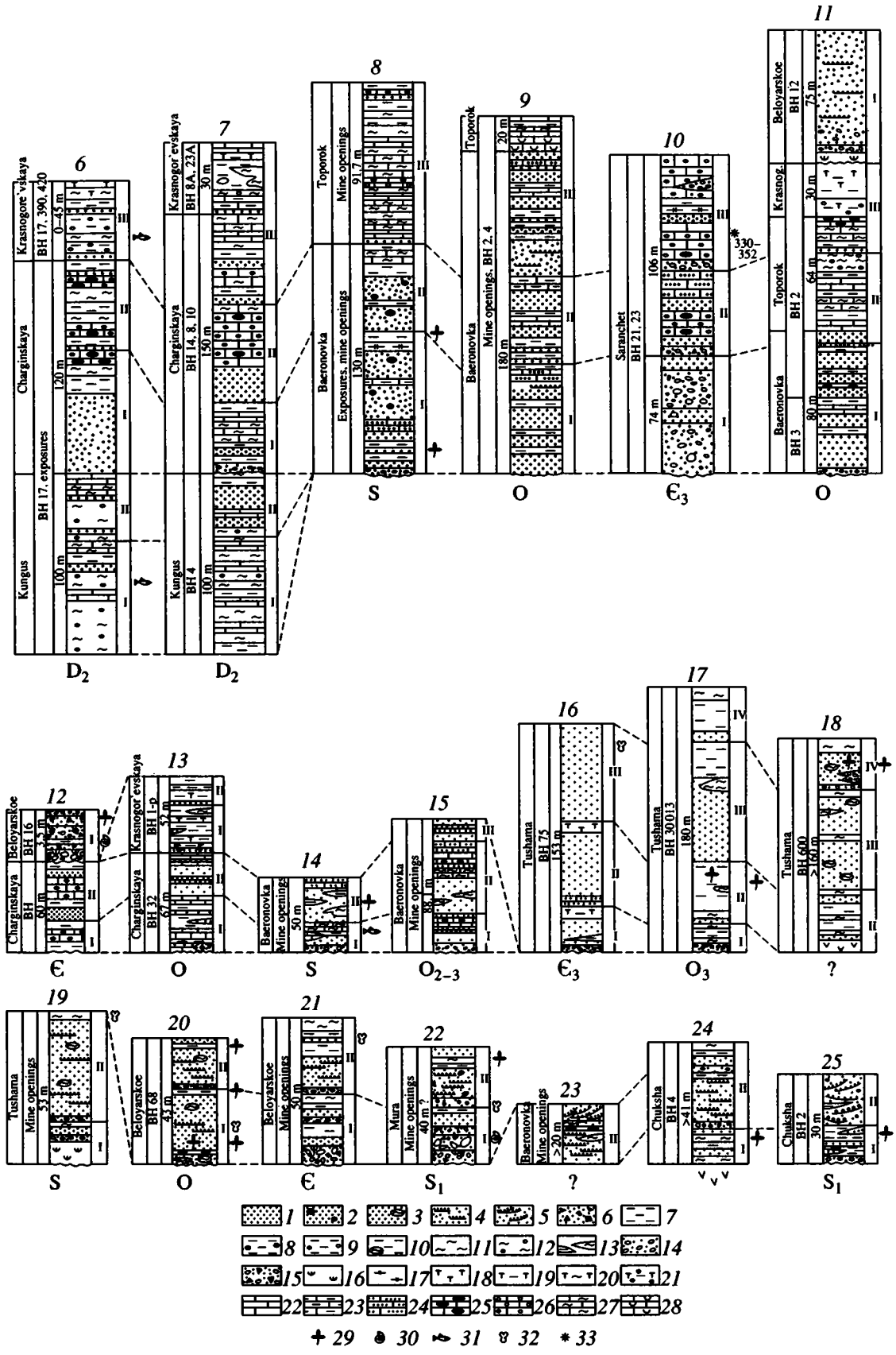
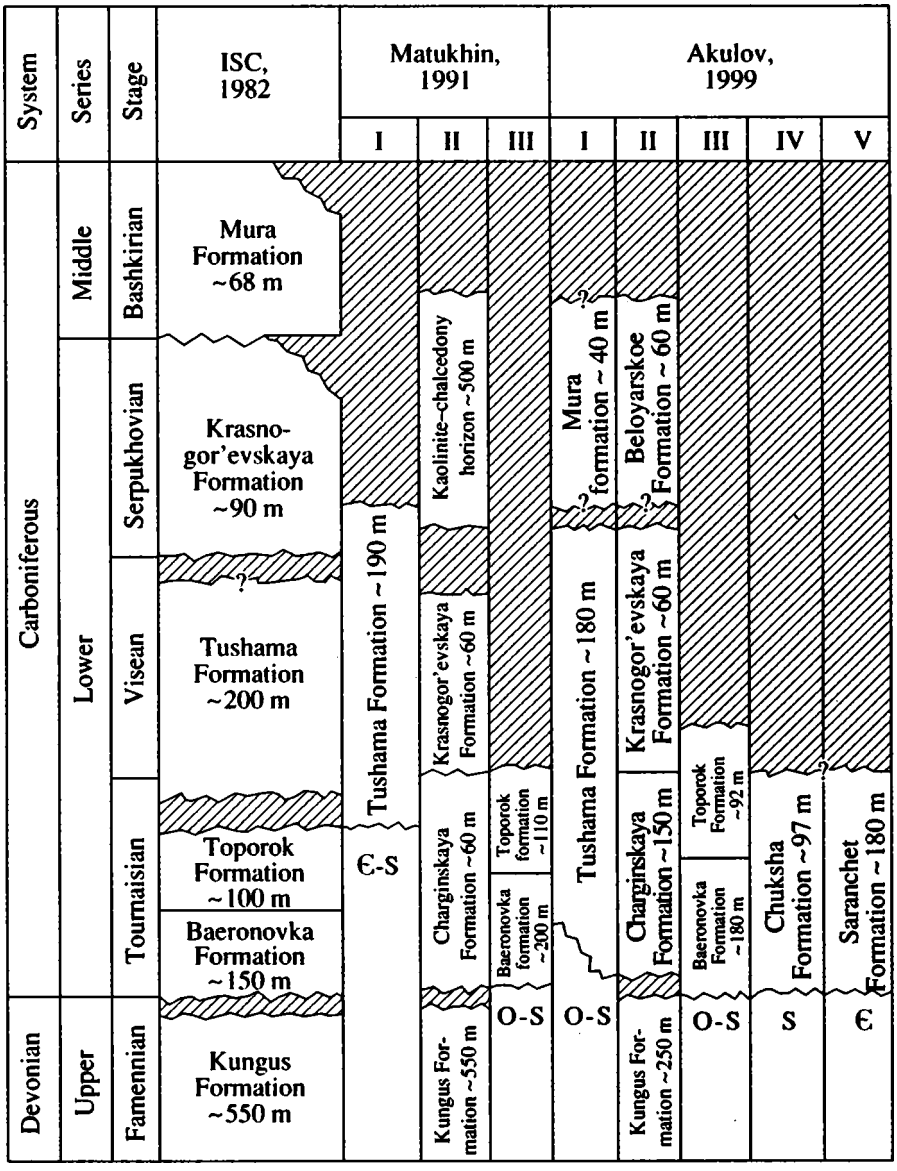

6
Krasnogorovskaya
Charginskaya
Kungus
BH 17, 390, 420
BH 17, exposures
0–45 m
120 m
100 m
D2
7
Krasnogorovskaya
Charginskaya
Kungus
BH 8A, 23A
BH 14, 8, 10
BH 4
30 m
150 m
100 m
D2
8
Toporok
Baeronovka
Mine openings
Exposures, mine openings
91.7 m
130 m
S
9
Toporok
Baeronovka
Mine openings, BH 2, 4
20 m
180 m
O
10
Saranchet
BH 21, 23
106 m
74 m
330–352
Є3
11
Beloyarskoe
Krasnogorovskaya
Toporok
Baeronovka
BH 12
BH 2
BH 3
75 m
30 m
64 m
80 m
O
12
Beloyarskoe
Charginskaya
BH 16
BH
Є
13
Krasnogorovskaya
Charginskaya
BH 1-p
BH 32
O
14
Baeronovka
Mine openings
50 m
S
15
Baeronovka
Mine openings
88.1 m
O2–3
16
Tushama
BH 75
153 m
Є3
17
Tushama
BH 30013
180 m
O3
18
Tushama
BH 600
160 m
?
19
Tushama
Mine openings
53 m
S
20
Beloyarskoe
BH 68
43 m
O
21
Beloyarskoe
Mine openings
50 m
Є
22
Mura
Mine openings
40 m ?
S1
23
Baeronovka
Mine openings
>20 m
?
24
Chuksha
BH 4
>41 m
25
Chuksha
BH 2
30 m
S1
1 2 3 4 5 6 7
8 9 10 11 12 13 14
15 16 17 18 19 20 21
22 23 24 25 26 27 28
29 30 31 32 33

Fig. 2. (Contd.)

General geochronological table for correlation of Late Devonian and Early Carboniferous sedimentary formations



I—Angara–Nizhnyaya Tunguska interfluve; II—Kan–Taseeva interfluve; III—Chuna-Biryusa interfluve; IV—Uda-Vikhorevka interfluve; V—Taishet region.

and dolomitic varieties dominate among siltstone. Their cement is argillaceous–dolomitic and argillaceous. Mudstone beds are of montmorillonite and, rarely, montmorillonite–kaolinite composition, with the admixture of vermiculite, chlorite, and quartz. Light gray conglomerate beds include pebbles (up to 5 cm in size) of quartz, jasper, chalcedony, and highly weathered siltstone and mudstone.

Garnet, zircon, tourmaline, sphene, and Cr-spinel dominate in a heavy fraction of the rocks.

These deposits are poor in paleontological remains. Their Early Carboniferous age has been established on the basis of single findings of coalified Lycopodiopsida trunks.

The Saranchet Formation (named after the synonymous village) was distinguished in 1959 by V.G. Komarevskii and I.V. Zhukov as corresponding to rocks exposed in Cliff Bychikha (the Tumanshet River middle course). They referred it to the Late Devonian and correlated with the Amonash Formation of the Rybinskii depression.

Deposits of the formation can clearly be subdivided into three mesocycles: terrigenous (up to 74 m), carbonate–terrigenous (46 m), and terrigenous–carbonate

(up to 60 m). Terrigenous deposits of the first mesocycle occur with a sharp angular unconformity on the erosion surface of the Upper Cambrian deposits and are mainly represented by conglomerate beds, which enclose separate boulders and lenses of gritstone and are intercalated with sandstone interlayers.

Carbonate deposits of the second and third mesocycles are mainly composed of limestone, calcareous sandstone, silicified limestone, and marl.

The formation rocks are widespread in the Bolshaya and Malaya Verblyuda, Tumanshet, Elanka, Slyudyanka, and Bychikha river basins. According to Komarevskii and Zhukov (1960), siltstones from a borehole drilled on the right slope of the Bychikha River valley (depth interval of 64.7–65.9 m) yielded spores, among which L.D. Gomulevskaya identified *Leiotriletes microrugosus* (Ibr.) Naum., *L. atavus* Naum., *Retusotriletes microthelis* Naum., *R. parvimammatus* Naum., *R. communis* Naum., *Hymenozonotriletes krestovnikovii* Naum., *Trachytriletes* sp., *Archaezonotriletes basilaris* Naum., *Stenozonotriletes conformis* Naum., *S. simplicissimus* Naum., *Acanthotriletes tenuispinosus* Naum., *A. squamosus* Naum., *Lophotriletes exiquis* Naum., *L. turbinatus* Naum. These taxa are characteristic of both the Upper Devonian and Lower Carboniferous deposits.

The U–Pb dating of limestone sample collected near the Settlement of Vengerka (the Tumanshet River middle course) yielded the age of 330 ± 30 Ma, whereas the Pb–Pb date characterizing it (with a large dispersion of isotope ratios) is 352 ± 66–69 Ma (Akulov and Fefelov, 1991). It is likely that the limestone was formed in the terminal Late Devonian–initial Early Carboniferous time, although this assumption should be verified.

As indicated earlier, the rock sequence of the Toporok and underlying Baeronovka formations is a stratigraphic analog of the Charginskaya Formation. This is evidenced by imprints of fossil flora and by their similarity in composition (whitish quartz sandstone at the base and silicified limestone interlayers at the top).

The Toporok Formation (named after the Toporok River, a tributary of Usolka River) has a rather remarkable history that is directly concerned with the name of M.M. Odintsov, who studied the Chuna–Biryusa watershed in 1932–1940 and was first to distinguish Cambrian and Silurian deposits there. He subdivided the upper part of the Silurian section into the Baeronovka, Toporok, and Karabula horizons (Odintsov *et al.*, 1937). Later on, Odintsov (1960) revised the age of that part of the section and referred the Baeronovka and Toporok horizons to the Upper Devonian, discarding the Karabula Horizon as a valid subdivision.

It should be noted that the boundary between the Baeronovka and Toporok formations is vague, established arbitrarily on the basis of a gradual increase in abundance of variegated rocks downward in the section, where sandstone interlayers are less frequent. Hence, the latter unit differs from the Baeronovka Formation. It is less rich in terrigenous material, has a higher carbonate content, and shows presence of silicified limestone interbeds of reference value, which are rather thick and persistent along the strike. At the same time, sandstone and siltstone are similar in both formations, which are poorly characterized in terms of paleontology. Tubular forms of organic origin were encountered in limestone and siltstone. V.P. Maslov and B.S. Sokolov attributed these forms to algal stromatolites. Their age has not been established, and none of other fossil remains has been found.

The Baeronovka Formation (after the Baeronovka River, a right tributary of Biryusa River),[1] as is mentioned above, was originally distinguished as the Silurian Baeronovka Horizon by Odintsov (1940). V.L. Kanitskii was first to describe it as a formation in 1961. Deposits of the formation are widespread in the Poima–Biryusa SSZ, where they occur with a stratigraphic unconformity on Lower Paleozoic rocks. The formation is represented by compact and poorly-cemented sandstone, siltstone, mudstone, marl, and conglomerate beds. Conglomerate beds, usually confined to the base of the formation, occur not everywhere. Their thickness does not exceed 0.3 m. These are grayish yellow, small-pebbled rocks. Pebbles consist of quartz, quartzite, siltstone, and sandstone. The section begins more often with grayish white sandstone beds, enclosing in places small, separate pebbles. Main rock types of the formation are fine- to medium-grained sandstones of yellowish to greenish and whitish gray color, and of quartz to feldspathic–quartz composition. They have carbonate, and more rarely, argillaceous or siliceous cement.

Siltstone and mudstone form thin interlayers, whereas marl and carbonate rocks occur in lenses that quickly pinch out. These massive and hard rocks are of gray, greenish gray, yellow, and white color.

In general, two to three mesocycles are distinguishable in the formation. Inconstant amount of mesocycles is related to a sharp decrease in the formation thickness when passing from sections of the southern (180 m) to northern (88.1 m) and eastern (50 m) parts of the Poima–Biryusa SSZ.

Kanitskii (1961) found plant imprints in siltstone of the Baeronovka Formation near the Village of Nizhnyaya Gogolevka. Among the imprints, N.M. Petrosyan identified that of *Lepidodendropsis theodorii* (Zal.) Jongm., known from Lower Carboniferous (lower Tournaisian) deposits). Later on, Kanitskii (1965) collected fossil plants from similar deposits north of Taishet. According to D.I. Ermolaev, the plants belong to primitive lepidophytes (Lycopsida). Fossil fish remains from deposits of the Baeronovka Formation were reported long ago. In 1958, A.L. Kalinovskii

[1] In the volume on Carboniferous and Permian of the Stratigraphic Dictionary of the USSR (1997), it is stated that the formation is named after a tributary of the Usolka River, but this one has no tributary with such a name.

found fish scale in these deposits during his geological route along the right slope of the Uda River valley, between its junction points with the Anga and Sukhaya Anga rivers. D.V. Obruchev established that scale belongs to a new acanthodian species (*Lomphodus* sp. nov.), the age of which falls within the Tournaisian Age of the Early Carboniferous.

Turning back to the Charginskaya Formation, we should point out that it is conformably overlain by the Krasnogor'evskaya Formation, deposits of which are widespread west of Novobiryusinsk.

The Krasnogor'evskaya Formation (after the Village of Krasnogor'evskaya) was first recognized among the Lower Carboniferous strata in 1955 by B.A. Fuks and I.M. Shakhanovskii.

A peculiar feature of the formation is the lack of limestones and presence of tuffs in its composition. The unit consists of monotonically alternating mudstone, sandstone, tuffite, and siltstone beds, the last two being prevailing. The light gray color with a bluish, greenish, more rarely yellowish tint is typical of the formation rocks.

Sandstones of quartz to quartz–feldspathic composition have a calcareous cement and show inequigranular or massive texture. Greenish gray tuffite beds are usually hard, often porcelaneous, with conchoidal fracture. Grumous siltstones of variable green tints include scattered small (up to 1 cm) pebble of green mudstone. Clayey siltstone may grade into mudstone. Ash-sized fragments of volcanic glass may represent up to15% siltstone composition. Glass fragments are of a peculiar hook-like, curved shape. They are usually replaced by radiated aggregates of yellowish-green chlorite. Mudstone minerals are chlorite, montmorillonite, and hydromica. In places, a distinct lamination is observed in rocks of the formation. It is formed by thin (1–2 mm) interlayers of clayey material in siltstone and by sandy laminae in mudstone. The thickness of the formation varies from 40 to 60 m.

In 1947, A.V. Aksarin found imprints of *Strepsodus siberiacus* Chab. (determination by D.V. Obruchev) at the depth of 273 m in borehole 266 drilled near the Village of Pereyaslavka. Later on, Lebedev collected fossil flora from upper mudstone beds of the formation, which are exposed near the Village of Voznesenka. Anan'ev and Mikhailova recognized the following species of the flora: *Zalesskyodendron sibiricum* Radcz., *Z. alternans* (Schm.) Radcz., *Protosalanus distans* (Chachl.) Radcz., *Lyginopteris* sp., *Praelepidodendron varium* Radcz., *Sublepidodendron alternans* (Schm.) An. et Mikh., and *S. distans* (Chachl.) An. et Mikh.

It is significant that numerous imprints of *Sublepidodendron alternans* (Schm.) An. et Mikh. indicate the Visean (Early Carboniferous) age of their host beds. It seems plausible that the upper part of the formation corresponds to the lowermost Serpukhovian (table).

Almost a complete stratigraphic analog of the Krasnogor'evskaya Formation is the Tushama Formation.

The Tushama Formation (after the Tushama River, a tributary of the Angara River) was distinguished by M.F. Neiburg in 1956. Deposits of the formation are widespread in the northeastern part of the Angara River region where they occur with a stratigraphic unconformity on the erosion surface of Lower Paleozoic rocks. The formation thickness can be as great as 180 m. Four mesocycles, clearly distinguishable in the unit, are well traceable from one section to another. Sandstone is a dominant rock of the formation; siltstone, mudstone, tuffite, tuffaceous shale, and tuffite represent 20% of its volume. The composition of the Tushama Formation is known in detail (Pavlov *et al.*, 1974, 1990). Its age is established on the basis of the fossil flora found.

In 1949, D.A. Tugolesov collected fossil flora at the Krasnyi Yar exposure, which later became the type section of the Tushama Formation. Neiburg (1956) who studied his collection determined the decorticated cortex of lepidophytes as *Knorria* and *Bergeria*, the age of which falls within the Early Carboniferous (Visean and Serpukhovian).

Flora found in the same exposure by M.I. Graizer and F.S. Ul'masvai allowed Anan'ev *et al.*, (1969) to attribute the Tushama Formation the late Tournaisian–early Serpukhovian interval.

In 1982, we collected a floral assemblage in the vicinity of the Boguchansk Power Station (the key section 17). It consists of *Sublepidodendron neuburgae* Anan. only. Plant imprints were also found in a borehole drilled near the Tushama River mouth. Imprints of thin trunks with large cushions, which belong to *Caenodendron sibiricum* (Kovb.) Radcz., were encountered in gray sandstone at a depth of 189 m. Flora remains from the borehole made it possible to refer enclosing rocks to the interval from the late Tournaisian to the early Serpukhovian included.

The composite Lower Carboniferous section is crowned by a kaolinite–chalcedony horizon that rests on erosion surface of the Krasnogor'evskaya and Charginskaya formations, and on older rocks, those of the Archean included (Matukhin, 1991). According to ISC resolutions (*Resheniya...*, 1982), the kaolinite–chalcedony horizon, which was previously attributed (along with younger coal-bearing deposits) to the Tushama Formation, is named the Mura Formation and referred to the Middle Carboniferous that is quite unjustified in our opinion.

It would be valid for the isolated Kulakovo–Pogromnaya and Kokui depressions of the Angara River region and only for their upper coal-bearing Tushama Subformation. According to G.N. Troshkova, (reference in Pavlov, 1974), conglomerates of the lower Tushama Subformation occurring in these depressions are of the Serpukhovian age and concurrent to the kaolinite–chalcedony horizon of the Kan–Taseeva SSZ, whereas the upper Tushama Subformation corresponds to the Mazur Formation of the Kuznetsk Basin (Middle

Carboniferous) or to the lower part of the Kat Formation (Middle–Upper Carboniferous).

The fact that rocks of the kaolinite–chalcedony horizon are widespread in the Kan–Taseeva SSZ, and a reliable palynological dating of the horizon (reference to data of L.N. Peterson in Yuferev *et al.*, 1980) allowed us to distinguish it as an independent Beloyarskoe Formation (after synonymous village in the Kansk district).

The Beloyarskoe Formation has its 200-m-thick type section exposed at the right bank of Kan River (3 km downstream from the Village of Beloyarskoe). Conglomeratic breccias of the unit include well-rounded (variegated) pebbles and debris of quartz, jasper, agate, quartzite, and individual boulders of poorly rounded chalcedony. The rocks are ochreous almost everywhere and carbonatized in places. Their roof (about 0.8 m) shows cuirass-like textures and the uppermost part (up to 0.15 m) represents a true cuirass. Matrix of the rocks (up 50–70%) is composed of angular quartz and chalcedony grains cemented by kaolinite or siderite.

A distinct graded bedding is typical of the section, in which the basal conglomeratic breccia grades upward into gritstone and then to sandstone that includes isolated angular fragments of chalcedony. Above, we see the new rhythm beginning again with conglomeratic breccia, and the graded structure resumes, though the rocks fragments are less roundness here till the appearance of sedimentary breccia.

In its stratotype, the Beloyarskoe Formation rests on the eluvial weathering crust above limestone of the Charginskaya Formation. There are areas where conglomeratic breccia fills in karst depressions in limestone (Pavlov and Budnikova, 1977). In places, white to cream-colored, brecciated and dense clayey rock with cherty concretions occurs in small relief depressions at the base of the formation (Akulov, 1984). The bed is from 2 to 5 m thick and overlain by a layer (0.2–5 m thick) of white and dense kaolin mudstone with cherty inclusions, which are similar in composition to flint clay (an exposure on Elan' River, 1.4 km upstream of the mouth).

Kaolinite–chalcedony deposits of the Beloyarskoe Formation also occur as isolated "spots" of 4 km^2 in size at the foot of the Yenisei Range, where they overlie with a sharp angular unconformity the Archean biotite granite of the Tarak complex. The formation section is uniform in general; its main element is the conglomeratic chalcedony–kaolin breccia (up to 20 m thick) occurring at the base. This rock is composed of redeposited products of the old weathering crust: light gray kaolin clay, variegated cavernous jasper, chalcedony, quartzite, and well-rounded pebbles of underlying rocks. The sequence is strongly ocherous, enclosing lenses and interlayers of yellow, inequigranular sandstone with rare quartz pebbles. Overlying beds consist solely of sandstones with thin siltstone interlayers. These quartz to feldspathic–quartz sandstones are light gray, fine- to coarse-grained, composed of semirounded and rounded grains of quartz (60–80%), feldspar (10–20%), siliceous rocks, and muscovite. Their cement is basal, argillaceous (kaolinite) or carbonate–argillaceous.

It should be pointed out that lithology of the formation is not persistent. In some areas, conglomeratic breccia dominates, whereas in others, sandstone and conglomerate prevail. Therefore, it is difficult sometimes to establish the upper boundary of the formation. If it is the case, the boundary is arbitrarily placed at the level of disappearance or sharp reduction of the coarse-clastic material, above which we see the Middle–Upper Carboniferous black coaly siltstones and mudstones with coal interlayers or coal seams.

According to Blagodatskii and Usaar (1968), A.S. Khamentovskii and A.A. Predtechenskii found poorly preserved imprints of pelecypods Modiolopsidae in rocks of the Beloyarskoe Formation at the Gryaznyi Creek (a left tributary of Bryakich River). In 1976, F.M. Gaintsev discovered in siltstones of the formation, at the right bank of Ezhik River, a monotypic assemblage that consists, according to determination of V.M. Kovbasina, of *Angarodendron obrutschevii* Zal. cortex. The species appeared in the Ostrog time (Serpukhovian Age) to become prosperous in the Mazurovo period (Bashkirian and Moscovian ages).

If we take into account fossils described above and numerous spore-and-pollen spectra described by G.N. Troshkova (Pavlov, 1974) and L.N. Peterson (Yuferev *et al.*, 1980), the Beloyarskoe Formation can be referred to the time interval from the Serpukhovian to the earliest Bashkirian (table). Hence, the Murma Formation appears to be unnecessary, because it can be subdivided into units with different paleontological characteristics. Its lower part (the conglomerate horizon) can be considered the Beloyarskoe Formation, whereas the upper one (coal-bearing) that yields the Middle Carboniferous spores and pollen can be placed at the base of the Kat Formation.

The Mura Formation, deposits of which are stripped by mine openings and penetrated by boreholes in the Mura River upper courses, represents a stratigraphic analog of the Beloyarskoe Formation (Fig. 1).

The Mura Formation (after the Mura River, a tributary of the Angara River) was originally distinguished by B.M. Vladimirov and L.N. Zveder as the Mura Sequence. In this unit, they found pyrope-bearing conglomeratic breccia and sandstone, which are exposed in a narrow belt (0.2–0.8 km wide and about 8 km long) and overlie the Lower Paleozoic deposits. The facies analysis showed that talus to proluvium and alluvium-type deposits of the Mura Formation accumulated in an old valley (Akulov, 1990).

Rocks of the Mura Formation occur with a stratigraphic unconformity on the erosion surface of Ordovician and Silurian deposits. The most representative sections of the formation are stripped by mine openings

in upper reaches of the Artel'nyi Creek. At this site, disintegrated and weathered red beds of the Bratsk Formation are overlain by the following units (from the base upward):

(1) Conglomeratic breccia with rounded to angular quartz, quartzite, chalcedony, jasper, and silicified limestone fragments ranging in size from 0.5 to 10 cm; separate boulders as large as 40 × 40 cm are also encountered. Angular fragments dominate over the rounded ones. The rock is intensely ocherous and carbonatized in places (1.5 m thick).

(2) Greenish gray conglomerate with interlayers and lenses of limonitized sandstone; well-rounded pebbles are composed of greenish blue mudstone, grayish white limestone, and jasper (fragments of the latter rock are poorly rounded). Carbonatized sandy cement includes numerous grains of violet pyrope (1.6 m thick).

(3) Yellowish gray coarse-grained quartz sandstone with horizontal bedding (1.5 m thick).

(4) Pinkish gray gritstone consisting of quartz, jasper and mudstone grains; cement of the rock is calcareous (0.5 m thick).

(5) Light gray coarse-grained quartz sandstone with horizontal bedding and abundant angular jasper grains is (2.5 m thick).

(6) Yellowish brown, fine- to medium-grained sandstone (0.8 m thick).

(7) Greenish gray mudstone (1.0 m thick).

(8) Medium-grained quartz sandstone (1.0 m thick).

(9) Yellowish gray mudstone (0.7 m thick).

(10) Fine-grained quartz sandstone with argillaceous cement (0.8 m thick).

(11) Gray mudstone with horizontal wavy lamination (0.7 m thick).

(12) Dark gray, quartz sandstone with inclusions of gravel and pebbly material (5.7 m thick).

(13) Yellowish brown fine-grained sandstone with horizontal lamination (3.5 m thick).

(14) Dark gray carbonaceous mudstone (1.5 m thick).

(15) Light gray fine-grained sandstone with poor plant imprints (3.0 m thick).

(16) Yellowish gray fine-grained sandstone (4.6 m thick).

(17) Dark gray carbonaceous mudstone (0.5 m thick).

(18) Dark gray, fine-grained sandstone with poorly expressed horizontal lamination (3.6 m thick).

(19) Dark gray siltstone alternating with mudstone (2.0 m thick).

(20) Gray fine-grained sandstone (3.0 m thick).

The total thickness of the composite section is 40.0 m.

The age of the Mura Formation is debatable. During a decade (1959–1969), it was thought to be the Late Devonian (data of 1958 by B.M. Vladimirov and L.N. Zveder,). In 1971, S.A. Kashik and M.M. Odintsova referred these rocks to the Early Jurassic on the basis of spore-and-pollen complexes from various horizons above conglomeratic breccia. Coalified plant remains 3 mm in diameter were found in the latter. The wood cell structure is of the *Araucaria* type: in places, scalariform and spiral thickenings can be observed. Small and destroyed cuticle fragments are found as sole remains in fine-grained sandstone of the formation. Fragments are very small and poorly preserved. A peculiar feature of the plant detritus of the Mura deposits is the mineralization of a part of cuticles and the coalification, to a considerable extent, of other cuticle remains not easily detectable.

L.V. Bashkirov and L.N. Zveder reported in 1961 on pyrope-bearing conglomerates, which yield abundant, probably, redeposited fragments of echinoderm spines (O.A. Pavlinov's collection, determinations by O.K. Smirnova).

It was impossible to macerate spores and pollen from deposits under consideration, because they are almost "barren." An exception is the dark gray siltstone occurring at the top of the Mura Formation and penetrated by Borehole 63 at the watershed of the Dzhelinda and Kamishi rivers. This rock yielded the following spore species: *Acanthotriletes larvatus* Lub., *A. lunatus* (Kustova) Naum., and *Retusotriletes ciliaris* Peterson. The second and third taxons are indicator species of the Serpukhovian age. Both are typical of the kaolinite–chalcedony horizon (Graizer *et al.*, 1980) and of the upper part of the Tushama Formation in the Angara River basin (Pavlov, 1974). We believe, hence, that sediments of the Mura Formation accumulated during the Serpukhovian and probably at the beginning of the Bashkirian.

Summing up consideration of stratigraphy and correlation of Upper Devonian and Lower Carboniferous rocks, we should briefly mention the key geologic–tectonic structure that controlled their formation. It was the Oka–Vikhorevka flexure of a submeridional orientation, which hampered progradation of the old Poima–Biryusa bay eastward (Fig. 1). The flexure is 600 km long, its limbs are displaced for a distance of 400 m, and the eastern one is thrown up (Zamaraev and Kuznetsov, 1961). The flexure steepness increases from the south to the north, and the flexure strata are plunging per 1 km distance for 6 m in the Oka River area, for 50 m near Bratsk, and for 150 m and more in the Kova River upper reaches. The Oka–Vikhorevka flexure is genetically related to faults in the basement, representing their reflection in the sedimentary cover, as well as in the Upper Devonian and Lower Carboniferous strata. At present, a chain of ranges, which are composed of Ordovician and Silurian deposits and reflect complex differentiated block movements in the basement, extends along the western limb of the flexure (Fig. 1).

As is established in the diamond-bearing province of Yakutia, the diamond pipes are confined to tectonic dislocations, i.e., to flexures and associated faults within the field of Ordovician rocks, and the Oka–Vikhorevka flexure is one of the cases. In the Poima–Biryusa SSZ, diamondiferous placer of the Chuksha River is situated in its eastern edge (Fig. 1). Moreover, pyrope, a satellite mineral of diamond, is found in rocks of the Baeronovka, Tushama, Chuksha, and Mura formations. Conglomeratic breccias and conglomerates of the Mura Formation are most remarkable among these rocks. They contain on the average about 30 signs of pyrope per a cradle slime sample of 20 l in volume. It should be noted that 20 diamond crystals weighing 145.8 mg have been found in fluvial deposits of the Kova River as well. All these observations imply a possibility of spatial relationship between the Oka–Vikhorevka flexure and primary sources of diamond in the southern part of the Siberian platform, as was indicated earlier by Zamaraev and Kuznetsov (1961).

CONCLUSIONS

(1) Lower Carboniferous deposits of the Angara River region represent a carbonate–terrigenous sequence, stratigraphic units of which are represented in the detailed correlation scheme of reference sections and in the general geochronological table.

(2) Deposits of the kaolinite–chalcedony horizon are distinguished as an individual Beloyarskoe Formation.

(3) Coal-bearing deposits of the Mura Formation (formerly, the upper Tushama Subformation) of the Kokui and Kulakovo–Pogromnaya depressions are attributed to the Kat Formation, and the underlying beds (the lower Tushama Subformation) are correlated with the Beloyarskoe Formation. The Murma Formation is therefore omitted from the correlation scheme.

(4) Rocks of the Mura Formation are correlated with deposits constituting the Beloyarskoe Formation and dated at the Early Carboniferous (Serpukhovian).

(5) The Oka–Vikhorevka flexure was the main Middle Paleozoic structure controlling the process of sedimentation in the southern part of the Siberian platform.

(6) The territory adjoining the Oka–Vikhorevka flexure is suspected to be the site of primary diamond sources.

Reviewers V.A. Aristov and A.S. Alekseev

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Late Albian Flora from the Vicinity of the Prokhladnoe Village (the Crimea)

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Abstract—Paleobotanical and palynological data on the same units seemingly lead to a controversial interpretation of the late Albian flora and vegetation. For instance, an upper part of the upper Albian *Hysteroceras orbignyi* Zone (the Prokhladnoe Village, the Crimea) reveals abundant impressions of Taxodiaceae shoots (genus *Geinitzia*) and ferns of the *Gleichenia* genus, which associate with rare representatives of Schizaeaceae and Dicksoniaceae. At the same time, bisaccate pollen of the Pinaceae, Podocarpaceae, and other families and that of the Schizaeaceae, Dicksoniaceae, Gleicheniaceae ferns dominate in the palynological spectrum of the same unit. So data, similar for ferns, are quite different for conifers: the prevalence of Taxodiaceae among imprints is not reflected in the palynological spectrum, where their pollen is found as single specimens. In turn, the dominance and diversity of bisaccate pollen in the palynological spectrum does not correspond to abundance of these plants imprints, as we found only two of them. The combination of data and the analyzed reasons of their inconsistency make it possible however to reconstruct reliably the late Albian vegetation. As is established, the seacoast in the studied area was covered with various ferns, among which Gleicheniaceae and Schizaeaceae representatives dominated. Abundant *Geinitzia* shoots found in the area suggest that the coastal band with ferns was not wide, and that coniferous microphyllous forests closely approached the sea. Palynological data indicate a proximity of hill slopes thickly populated by conifers producing the abundant bisaccate pollen. Taken together, these data imply that the relief was rugged and variable, and the climate was warm and humid.

Key words: Late Albian, the *Hysteroceras orbignyi* Zone, Crimea, fossil plants, palynological spectrum, palynological assemblage, flora, vegetation.

INTRODUCTION

The Albian was a specific period in the history of higher terrestrial plants because the angiosperms radiation started at that time. Unfortunately, there are only few corresponding floras reliably dated within the European part of the northern coast of the Tethys. More than a century ago, Albian floras were discovered in Portugal (Saporta, 1894), in the vicinity of Kanev, and in the Podolia region (Radkevich, 1895). The middle Albian flora has been recently found near Simferopol' in the Crimea (Stanislavskii and Kiselevich, 1986), while the late Albian flora was revealed near Kanev (Pimenova, 1939; Doludenko *et al.*, 1987, 1988, 1992), and near the Prokhladnoe Village in the Bakhchisarai region of the Crimea (Krasilov, 1984; Doludenko *et al.*, 1999). Krasilov found Cenomanian plant remains there as well. The late Albian floras from the Kanev and Prokhladnoe sites were investigated using modern techniques: the epidermal–cuticular analysis, observations under scanning microscope, and anatomic study of leaves and wood.

Since findings of late Albian plants are rare, we recollected samples from the Exposure A_4 studied by Krasilov (deposits of the *Hysteroceras orbignyi* Zone). Plant imprints are discovered in silty clays, and palynological spectrum is macerated from these rocks. This work is a result of the combined paleobotanical and palynological study aimed to get a deeper insight into the Albian flora and vegetation of the study region, since both methods supplement and verify each other.

Samples collected by Doludenko in 1988 are deposited at the Geological Institute RAS (Moscow), collection no. 4857. Plant impressions are pictured by A.I. Nazarov and M.G. Moiseeva who also prepared drawings of *Gleichenia zippei.*

DESCRIPTION OF THE SECTION

Near the Prokhladnoe Village, the upper Albian deposits fill in the deep Mangush depression. The synonymous sequence is represented by different terrigenous facies: from coarse unsorted conglomerates to fine silty clay. Because of a complicated structure, the Mangush sequence was studied time and again using geophysical, petrographic, paleontological methods, and purposeful drilling. Palynological assemblages and clay mineralogy were analyzed to reveal the formation

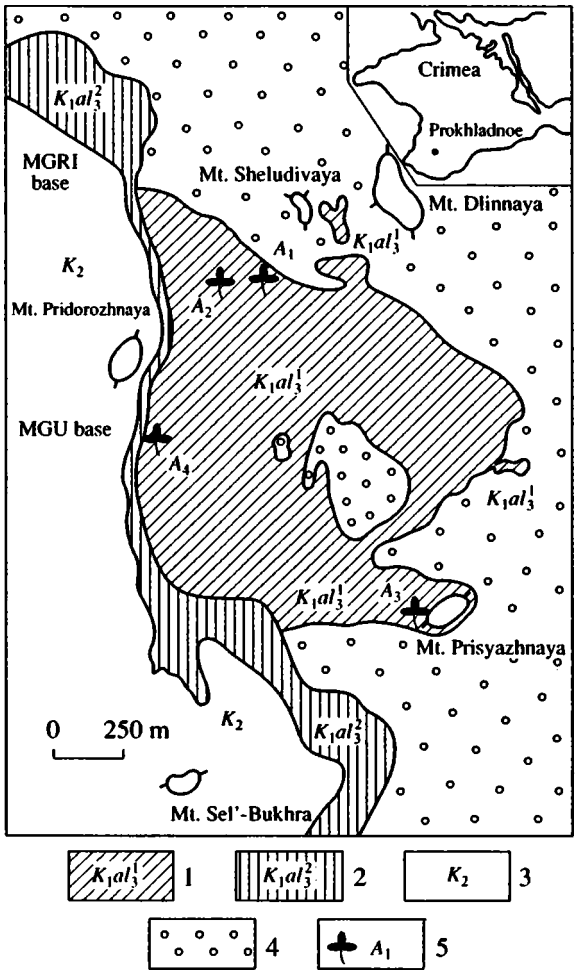


Fig. 1. Distribution map of upper Albian deposits near the Prokhladnoe Village, the Bakhchisarai region of the Crimea (after Yanin, 1976): (1) exposure area of clay, sandstone, and conglomerate of the *Hysteroceras orbignyi* Zone; (2) bedrock exposures of sandstone of the *Mortoniceras inflatum* Zone; (3) overlying Upper Cretaceous deposits; (4) pre-Albian deposits; (5) sampling sites A_1–A_4 of fossil plants, Krasilov (1984); Exposure A_4 is studied in this work.

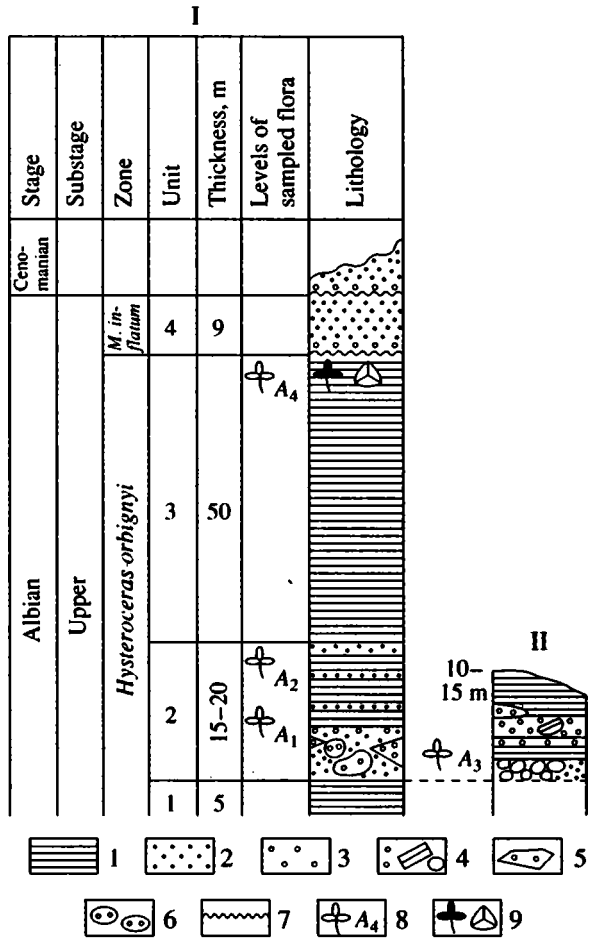


Fig. 2. Correlation scheme for flora-bearing sections of upper Albian deposits near the Prokhladnoe Village (after Yanin, 1964): (1) silty clay; (2) dense, well-cemented sandstone; (3) conglomerate; (4) boulder-and-lumpy conglomerate; (5) conglomerate lenses; (6) bun-shaped sandstone concretions with pebble; (7) erosion surface; (8) locality of flora found by Krasilov (1984); (9) sites of fossil plants and palynomorphs sampled by authors. Sections: (I) Mangushskii Creek– Mt. Pridorozhnaya, (II) Mt. Prisyazhnaya.

conditions of Mangush deposits (Yanin, 1964, 1976; Chernov and Yanin, 1975; Stafeev *et al.*, 1979, and others).

The Mangush depression is assumed to represent a floor of an ingression basin poorly connected with open-sea areas. Clastic material was transported to the depression from nearby and remote provenances. The sea ingression from southeast, from the Black Sea basin, was accompanied by transportation of size-variable rock fragments, a coarse material included. Fine material accumulated after the depression subsidence, when a stable channel for a sediment load was set up. The Mangush basin was of a flow-through type that is evident from euryhaline fauna that populated it.

A complete stratigraphic description of upper Albian flora-bearing deposits at the Prokhladnoe site was presented for the first time by Yanin (1964). He described them in a layer-by-layer manner and correlated the upper Albian deposits within the Kacha–Bodrak interfluve. A brief presentation of his data is given below.

In the Bakhchisarai region, upper Albian deposits are widespread from the Kacha River on the south to the Bodrak River on the north. They transgressively overlap underlying rocks, in places with the angular unconformity. In the vicinity of Prokhladnoe (Fig. 1), they overlie the Tavriya shale and underlie the basal small-pebbled conglomerate of Cenomanian glauconite sand and marl.

On the basis of lithology and faunal assemblages, upper Albian deposits are subdivided into the lower and upper parts corresponding to two upper Albian zones

Table 1. Composition of the late Albian flora from the Prokhladnoe Village region (Exposure A_4)

Species	Exposure A_4		Amount of specimens	Percentage		
	Krasilov, 1984	this work				
Lycopodites sp.	2		2	1.6	Lycopodiophyta	
Lycostrobus sp.	1		1			
Equisetum sp.	1		1	0.5	Equisetophyta	
Ruffordia goeppertii (Dunk.) Sew.	1	4	5	3.2	Schizaeaceae	Polypodiophyta
Anemia dicksoniana (Heer) Krassil.	1		1			
Osmunda sp.		5	5	2.7	Osmundaceae	
Sphenopteris sp.		2	2	1	Cyatheaceae	
Gleichenia zippei (Corda) Heer	22	28	50	27	Gleicheniaceae	
Sagenopteris variabilis (Vel.) Vel.	4		4	2	Caytoniales	
Pityophyllum sp.	2		2	1	Pinaceae	Pinophyta
Geinitzia cretacea Unger	86	16	102	57	Taxodiaceae	
G. reichenbachii (Geinitz) Sew.		4	4			
Rogersia angustifolia Font.	2		2	3.8	Angiospermae	
Sapindopsis variabilis Font.	3		3			
Aryskumia sp.	2		2			
Total	127	59	186			

(Fig. 2). The distribution of the *Hysteroceras orbignyi* Zone deposits is confined to the Prokhladnoe Village and adjacent areas (Mt. Prisyazhnaya, the road Prokhladnoe–Nauchnyi). The deposits fill in the Mangush deep erosion trough.

The zone section (Fig. 2, I) is divided into three units (Yanin, 1964, Fig. 2). Clays, 5–10 m thick, occur in the basal part (Fig. 2, Unit 1). They are overlain by Unit 2 of sandstone, conglomerate, and clay exposed on southern slopes of the Mts. Sheludivaya and Dlinnaya, in the Prokhladnoe Village, and in the vicinity of Mt. Prisyazhnaya. In the lower part of Unit 2, there is yellowish brown to variegated, polymictic, inequigranular, loose, cross-bedded, highly ferruginate sandstone that encloses ban-shaped concretions of dense sandstone and interlayers of coarse-pebbled conglomerate. In its middle and upper parts, the unit is composed of gray, yellowish-gray, brownish, and spotted, highly sandy clays with plant remains (Exposure A_1 after Krasilov, 1984). An interlayer of grayish-brown sandstone and small-pebbled conglomerate, which contain plant imprints (Exposure A_2 after Krasilov, 1984), occur at the top of clays. The total thickness of the unit near the Prokhladnoe Village is 15–20 m.

Coeval deposits at the southern flank of the trough (slopes and top of Mt. Prisyazhnaya and of the road loop area) are of a quite different structure. They correspond to a member of lumps and boulders, which encloses interlayers of small-pebbled conglomerates, highly ferruginate sandstone, and sandy clay with plant remains (Exposure A_3 after Krasilov, 1984, according to Yanin's opinion). The total thickness of Unit 2 near the Mt. Prisyazhnaya is up 10–15 m.

Higher in the section (Fig. 2, Unit 3), there are clays (50 m), the upper part of which is exposed only. The middle and upper parts of the unit are unexposed, concealed under gardens and buildings (Yanin, 1964). These gray, grayish brown, yellowish gray, or variegated clays are calcareous and silty. We collected abundant plant remains from the upper part of Unit 3 (Exposure A_4, Krasilov 1984) and sampled rocks of this interval for the palynological analysis. The interval also yielded echinoid remains *Holaster* sp. (determination by A.N. Solov'ev, Paleontological Institute, RAS) and thin shells of bivalves. Flora-bearing deposits underlie basal conglomerate and sandstone of the *Mortoniceras inflatum* Zone (Fig. 2, Unit 4). The site with flora is situated not far from the MGU training base, in a scour at the opposite side of the road about 25 m below the road bed.

CHARACTERISTICS OF THE FLORA

Table 1 presents a list of species from the Exposure A_4, which are identified by Krasilov (1984) and by us, amount of specimens and their percentage. As is evident, conifers (58%) and ferns (34%) prevail in the flora

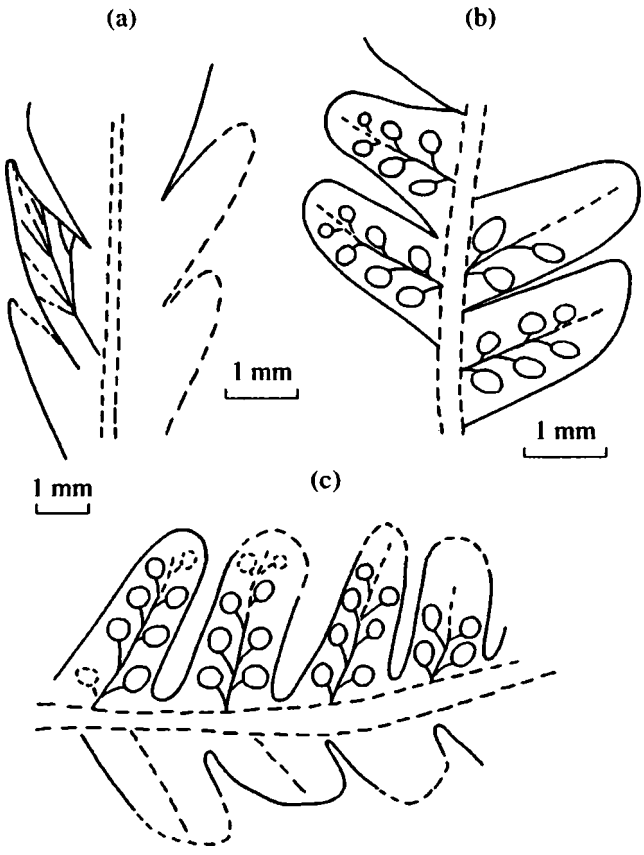


Fig. 3. Fern species *Gleichenia zippei* (Corda) Heer: (a) sterile pinna; (b, c) fertile pinnae with soruses.

composition. Dominants are conifers of Taxodiaceae family (*Geinitzia cretacea* and *G. reichenbachii*, 57%) and ferns of Gleicheniaceae family (*Gleichenia zippei*, 27%) (Plates I and II). The latter species is represented by both the sterile and fertile pinnae with soruses (Fig. 3), though the preservation of imprints was so poor that we failed to study the structure of sporangia and to extract spores. As the fertile pinnae of *G. zippei* were found for the first time in Albian deposits of the Ukraine, we present below a brief description of the species.

Fern *Gleichenia zippei* (Corda) Heer was described for the first time from Cretaceous deposits (shales) of Bohemia, Czech Republic (Corda in Reuss, 1846, p. 95, Plate XLIX, Fig. 2). In Russia, the species was described by Heer (1868), Vakhrameev (1952), Krishtofovich and Baikovskaya (1960), and by Krasilov (1967). We found 28 specimens in our collection, including two fertile and twenty six sterile (Plate I). Sterile fronds are bipinnate. The main rachis is 1.7 mm thick in the lower part of the frond and 0.3 mm thick in the upper part. Pinnae are linear–lanceolate, 1.2–3 cm long, 6–7 mm wide, attached in a successive order at the angle of 65°–75°. Thin (0.1–0.5 mm) shafts of pinnae are slightly curved towards the frond top. Pinnae are of the structure is catadromous. Pinnules are attached with the whole base at an angle. They are entire and different in shape, mainly tongue-shaped, falcate, and orbicular–triangular. The apex is narrowed and rounded. Pinnules are from 2.1 to 3.5 mm long and 1.4–1.8 mm wide at the base. They show pinnate venation (Fig. 3a) with the median rib slightly curved because of the lateral ribs branching. Lateral ribs (usually 3–4 from each side) dichotomize once in the rib middle part.

Fertile fronds are of the same structure as sterile. The median rib is clearly visible only at the base and in the middle part of pinnules, vanishing toward the apex. Soruses are arranged in two rows, by 3–4 in each, on both sides of the median rib (Fig. 3b, 3c). Fructification organs are not preserved, and only imprints of soruses are visible in the middle of lateral ribs; further bifurcation of lateral ribs cannot be traced. Soruses range in size from 0.2–0.3 mm in the lower part of pinnules to 0.1 mm at the apex.

The described fern species *Gleichenia zippei* (Corda) Heer differs from similar *G. porsildii* Seward and *G. gieseckianus* (Heer) Seward by lesser dimensions. Moreover, the shape of pinnules is less elongated as compared to *G. porsildii* Seward and not as curved as in *G. gieseckianus* (Heer) Seward (Seward, 1926).

G. zippei is widespread in Lower and Upper Cretaceous deposits of the Crimea, Western Europe, Greenland, western Kazakhstan, southern Primor'e, Sakhalin, and North America.

In addition to Gleicheniaceae, we distinguished ferns representing families Schizaeaceae (*Ruffordia goeppertii, Anemia dicksoniana*, 3.2%), Osmundaceae (*Osmunda* sp., 2.7%), and Cyatheaceae–Dicksoniaceae (*Coniopteris*, 1%).

Besides Taxodiaceae, rare representatives of Pinaceae (*Pityophyllum* sp., 1%) were found among conifers. Representatives of Lycopodiales (1.6%), Equisetaceae (0.5%), Caytheaceae (2%) are of minor importance. Percentage of angiosperms is 3.8%.

Plate I. Ferns from the lower zone of upper Albian deposits near the Prokhladnoe Village (the exposure below the road bed opposite the MGU base).

All specimens are deposited at the Geological Institute RAS, Moscow, collection no. 4857.

(1–5) *Gleichenia zippei* (Corda) Seward: (1) Sample 4857/1, imprint of the upper part of a frond (a), ditto (b) enlarged (4), counter-impression (c); (2) Sample 4857/3, a frond fragment; (3) Sample 4857/4, impression of the lower frond part and echinoid *Holaster* sp. (a), ditto (b) enlarged (2); (4) Sample 4857/2: imprint of fertile pinnae with soruses(a); ditto (b) enlarged (3); (5) Sample 4857/11, isolated pinna with soruses (3); (6, 7) *Ruffordia* sp.: (6) Sample 4857/14, a frond fragment (2); (7) Sample 4857/15 (2); (8, 9) *Osmunda* sp.: (8) Sample 4857/18, a frond fragment (2); (9) Sample 4857/6, upper parts of pinnae (2).

Plate I

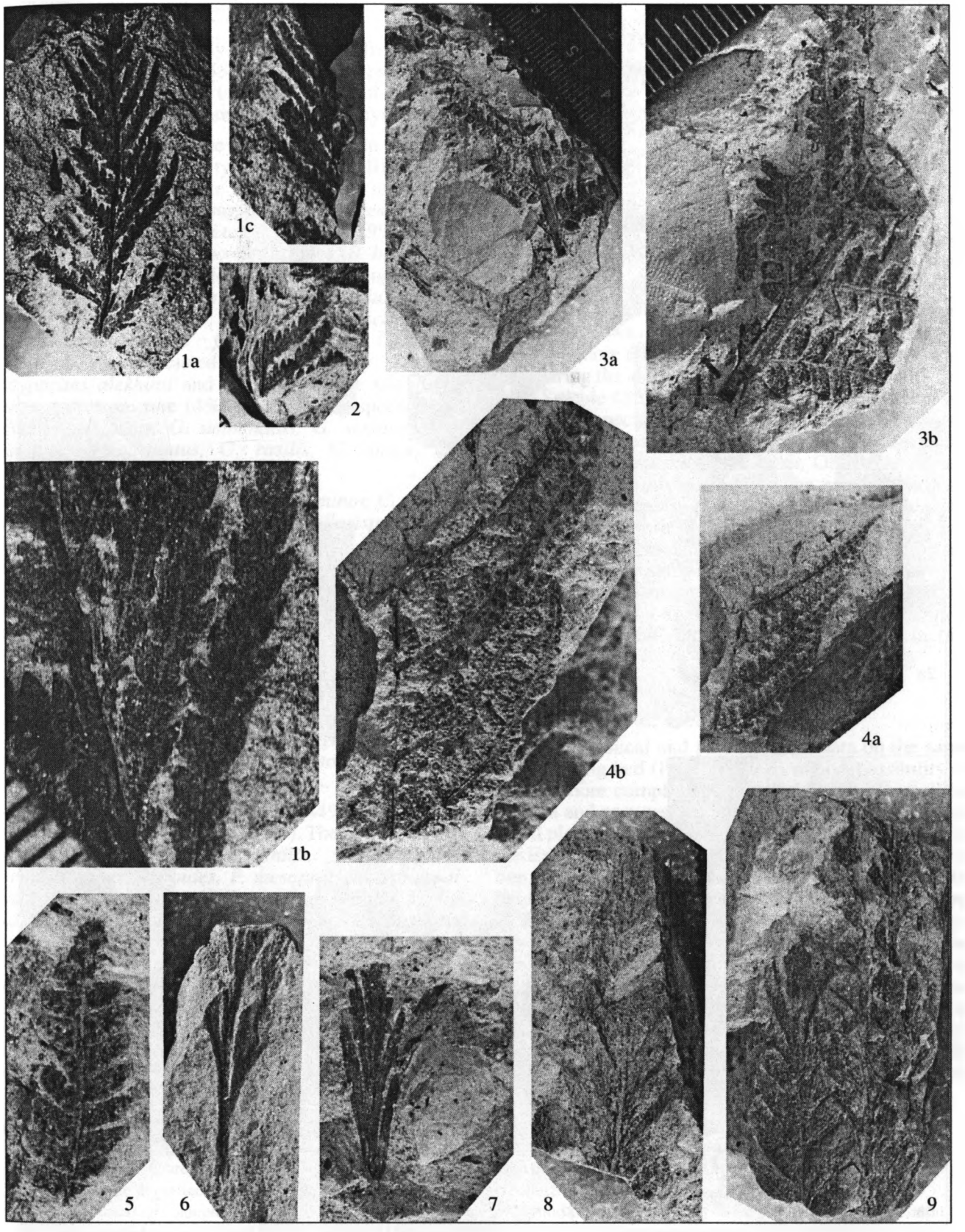
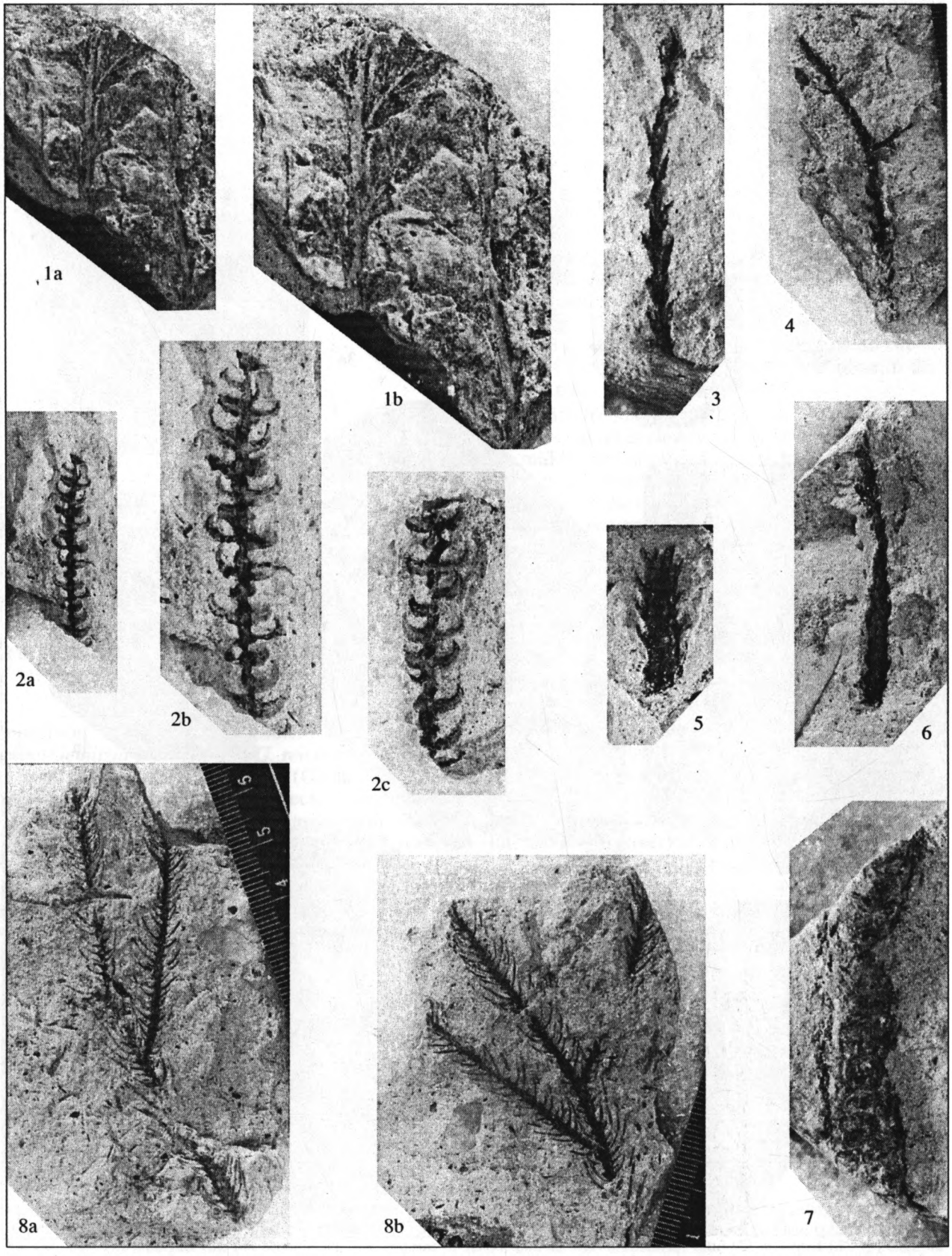


Plate II



CHARACTERISTICS OF PALYNOLOGICAL SPECTRUM

The microscopic study of macerates from upper Albian siltstone (Sample 4857/60) revealed presence of abundant organic matter: large fragments of plant tissue, spores, pollen, and rare dinoflagellate cysts.

The composition of the spore-and-pollen spectrum is noted for a great diversity (Table 2, Plate III), mainly due to spores, which make up 54 % of the total abundance of miospores. Among these are spores of ferns, especially of the Schizaeaceae family (19%) represented mainly by *Bikolisporites* species (*B. baconicus, B. toratus, B. transdanubicus*). Present as single specimens are *Klukisporites* sp., *Trilites triangulus, Appendicisporites concentricus, A. jansonii, A. tricornitatus, Costatoperforosporites fistulosus*, and *Cicatricosisporites* sp. associated with single-ray *Corniculatisporites alekhinii* and *C. striatus*. The Gleicheniaceae spores are rare (4%), representing species *Gleicheniidites laetus, G. umbonatus, G. senonicus, G. radiatus, G. carinatus, G. rasilis, G. minor*, and *Clavifera triplex*.

Among other spores, *Cyathidites minor, C. australis*, and *C. rarus* make up 9.5% and *Collarisporites fuscus*, 6%. Other spore taxa amount to 0.5–3%. Among these, there are spores of lycopsids (*Retitriletes dentimuratus, Retitriletes* sp., *Vadascisporites gregussii, Vadascisporites* sp., and *Densoisporites velatus*, 3% in sum). Spores of Osmundaceae (*Todisporites major*) and Matoniaceae (*Phlebopterisporites harskutensis*) are rare (0.5%). Among spores of the uncertain taxonomic position, there are *Leiotriletes* spp., *Duplexisporites generalis, Conbaculatisporites cretaceus, Conbaculatisporites* spp., *Distaltriangulisporites perplexus, Vinculisporites flexus, Laevigatosporites ovalis, Tripartina variabilis*, and many others.

The bisaccate pollen of conifers (44%) dominates in the spectrum of angiosperms (45%). The corresponding taxa are *Pinuspollenites similis, P. divulgatus, Piceapollenites exilioides, P. mesophyticus, Abiespollenites* sp., cedar pollen (*Parvisaccites radiatus, P. rugulatus*), *Podocarpidites luteus, P. multesimus*, and problematic *Disaccites* forms (12.5%). By the way, bisaccate pollen is produced not only by conifers but by other plants as well, for instance, by pteridosperms. *Caytonipollenites pallidus, Araucariacites australis, Taxodiaceaepollenites* sp. are found as single specimens. Pollen of *Cycadopites* and *Classopollis* has not been encountered.

Only one pollen grain of gymnosperm *Crassipollis vraconicus* Juh. and single dinoflagellate cysts *Gonyaulacysta* sp. were found.

Spores and pollen from deposits of the *Hysteroceras orbignyi* Zone at the Prokhladnoe site were insistently studied by Smirnova. As a result of her and other palynological investigations, a brief characteristics of the spore-and-pollen assemblage was published (Kuvaeva, 1966; Kuvaeva and Yanin, 1973; Smirnova, 1997; Stafeev *et al.*, 1977), a new genus *Corniculatisporites* Kuvaeva, 1972, new species of this genus (Kuvaeva, 1972; Smirnova and Shamma, 1987), and genus *Gregussisporites* Juhasz et Smirnova, 1985, were described.

Recently, a comprehensive description of miospores composition was published by Smirnova (1999) who analyzed more than 20 spectra characterizing this zone. Comparing the assemblage studied previously with that from Sample 4857/60, we found an undeniable similarity of their species composition. Among some distinctions, the principal one is an insignificant amount of the Gleicheniaceae spores in the latter. Of all the spectra studied before, only one shows the same abundance rate (8%), whereas percentage of Gleicheniaceae spores in others varies from 18 to 32%. In addition, fern spores are less abundant in Sample 4857/60 (54 versus 75–92% in other samples). By contrast, this sample is more rich in bisaccate pollen of conifers (38%), and proportions of *Pinuspollenites, Piceapollenites, Parvisaccites*, and *Podocarpidites* are higher in this group.

DISCUSSION

Paleobotanical and palynological data on the same flora-bearing bed (Fig. 2, Unit 3) offer a possibility to outline more completely and reliably the composition of flora and past vegetation. Comparing percentages of fossil plant species and main spore and pollen groups in the Exposure A_4 near the MGU training base (Table 3), one can see that ferns and conifers dominate in both cases. But if Gleicheniaceae ferns prevail, making up 27%, in plant imprints, among which Schizaeaceae represent only 3%, Osmundaceae 2.7%, and Cyatheaceae 1%, the leading role among spores belongs, according to palynological data, to Schizaeaceae (19%) and Cyatheaceae (9.5%), whereas Gleicheniaceae make up 4%, Osmundaceae 3%, and Matoniaceae 0.5%. Moreover, a considerable amount of spores of an uncertain taxonomic position (14.5%) was found. As is noted

Plate II. Ferns and conifers from the lower zone of upper Albian deposits near the Prokhladnoe Village (all samples, except for that pictured as figure 8, are from the exposure below the road bed opposite the MGU base).

All samples are deposited at the Geological Institute RAS, Moscow, collection no. 4857/60.

(1) *Sphenopteris* sp.: Sample 4857/17: a pinna fragment (a) enlarged (2), ditto (b) enlarged (3); (2) *Geinitzia reichenbachii* (Geinitz) Seward, Sample 4857/20: a shoot fragment (a), ditto (b) enlarged (2), counter-impression (c) enlarged (2); (3–8) *Geinitzia cretacea* Unger: (3) Sample 4857/29, a shoot fragment, (4) Sample 4857/24, (5) sample 4857/27, (6) Sample 4857/26, (7) Sample 4857/25, (8) Sample 4857/22: a shoot fragment (a) and counter-impression (b).

Table 2. Spore and pollen taxa from the lower zone of the upper Albian, the Prokhladnoe Village site (Exposure A_4 facing the MGU training base, below the road bed)

Spores and pollen	Content, %
Order Bryophyta	
Phaeocerosporites purus (Deak, 1964) Juhasz, 1980	0.5
Order Lycopodiophyta	
Vadascisporites gregussi Juhasz, 1975	1
Vadascisporites sp.	+
Retitriletes dentimuratus (Brenner, 1963) Juhasz, 1975	+
Retitriletes sp.	+
Densoisporites velatus Weyland et Krieger, 1953	2
Order Polypodiophyta	
Family Osmundaceae	
Conbaculatisporites cretaceus Deak, 1964	1
Conbaculatisporites sp.	1.5
Todisporites major Couper, 1958	0.5
Family Schizaeaceae	
Klukisporites sp.	+
Trilites triangulus Kedves, 1964	3
T. minor (Juhasz, 1972) S.K. Srivastata, 1975	+
Bicolisporites baconicus (Juhasz, 1972) Juhasz, 1977	6.5
B. toratus (Weyland et Greifeldt, 1953) S.K. Srivastava, 1975	6.5
B. transdanubicus (Juhasz, 1972) Juhasz, 1977	2.5
Acritosporites kyrtomis Juhasz, 1979	+
Cicatricosisporites venustus Deak, 1964	+
Cicatricosisporites sp.	+
Appendicisporites concentricus Kemp, 1970	+
A. tricornitatus Weyland et Krieger, 1953	+
A. jansonii Pocock, 1964	0.5
Costatoperforosporites fistulosus Deak, 1962	+
Corniculatisporites alekhinii (Bolchovitina, 1953) Kuvaeva, 1972	+
C. striatus (Deak, 1963) Kuvaeva, 1972	+
Family Gleicheniaceae	
Gleicheniidites senonicus Ross, 1949	0.5
G. laetus (Bolchovitina, 1953) Bolchovitina, 1968	2.5
G. umbonatus (Bolchovitina, 1953) Bolchovitina, 1968	0.5
G. radiatus (Bolchovitina, 1953) Bolchovitina, 1968	0.5
G. rasilis (Bolchovitina, 1953) Bolchovitina, 1968	+
G. minor Döring, 1965	+
Ornamentifera granulata (Grigorjeva, 1961) Bolchovitina, 1968	+
Family Matoniaceae	
Phlebopterisporites harscutensis Juhasz, 1979	0.5
Family Cyathaceae–Dicksoniaceae	
Cyathidites minor Couper, 1953	6.5
C. australis Couper, 1953	1
C. rarus (Bolchovitina) Deak	2

Table 2. (Contd.)

Spores and pollen	Content, %
Sporae Incertae Sedis	
Duplexisporites generalis Deak, 1962	2.5
Vinculisporites flexus Deak, 1964	+
Distaltriangulisporites perplexus (Singh, 1964) Singh, 1971	+
Collarisporites fuscus Deak, 1964	6.5
Leiotriletes microrugosus Naumova, 1953	5.5
Laevigatosporites ovatus Wilson et Webster, 1946	+
Order Pinophyta	
Caytonipollenites pallidus (Reissinger, 1938) Couper, 1958	+
Podocarpidites multesimus (Bolchovitina, 1956) M. Petrosjanz, 1971	3
P. luteus (Bolchovitina, 1956) M. Petrosjanz, 1971	1.5
Pinuspollenites similis (Balme, 1957) M. Petrosjanz, 1971	2.5
P. minimus Kemp, 1970	1
P. divulgatus (Bolchovitina, 1956) M. Petrosjanz, 1971	6
Piceapollenites exilioides (Bolchovitina, 1956) M. Petrosjanz, 1971	5.5
P. mesophyticus (Bolchovitina, 1956) M. Petrosjanz, 1971	8.5
Abiespollenites sp.	2
Parvisaccites radiatus Couper, 1953	1.5
P. rugulatus Brenner, 1963	+
Disaccites	12.5
Araucariacites australis Couper, 1958	0.5
Taxodiaceaepollenites sp.	0.5
Order Magnoliophyta	
Crassipollis vraconicus Juhasz, 1979	0.5
Order Dinophyta	
Gonyaulacysta sp.	0.5

above, however, the abundance of Gleicheniaceae is higher (18–32%) in palynological spectra characterizing deposits of the *Hysteroceras orbignyi* Zone in other localities.

In contrast, there is a complete discrepancy between macro- and micropaleontological data characterizing conifers. Taxodiaceae dominate in plant imprints (57 versus 58% of total amount of conifers), whereas pollen of *Pinuspollenites, Piceapollenites, Podocarpidites*, and others prevails in the palynological spectrum (Table 3), where pollen of Taxodiaceae (and of Araucariaceae) is represented by single specimens (0.5%). What is the reason?

The bisaccate pollen of conifers (Plate III, 17–20) prevails in the palynological spectrum owing to its volatility that is favorable for dispersal over a large distance from forests, which covered elevated areas. In this instance, the forests were likely growing far from the coast of a sea bay, since only two leaf imprints were found (leaves of *Pityophyllum* sp.). The rarity of the Taxodiaceae pollen specimens seems to reflect the fact that these plants produce much less pollen as compared, for instance, to Pinaceae, and their pollen is of a low resistance against fossilization. This may be also connected with two other circumstances: conifers of genus *Geinitzia* producing that pollen could grow at some distance away from the coast, and the pollen (Plate III, 22) could not be transported for a considerable distance owing to its morphological structure. It should be pointed out that *Geinitzia* shoots are mostly preserved as small fragments (Plate II, 3–7), though a large fragment (Plate II, 8) was also found near the Prokhladnoe Village in another exposure. Such a fragmentation of *Geinitzia* twigs (and other plant material) could be related either to the action of alongshore currents and bioturbation (Krasilov, 1984), or to their transportation from the place of growing to the site of burial. Remains of *Gleichenia zippei* are preserved much better (Plate I, 1, 3, 4), as this fern likely grew closer to the burial site, i.e., directly on the bay coast.

We realize of course that the considered ratios and percentages are conventional, since we compare the

Plate III

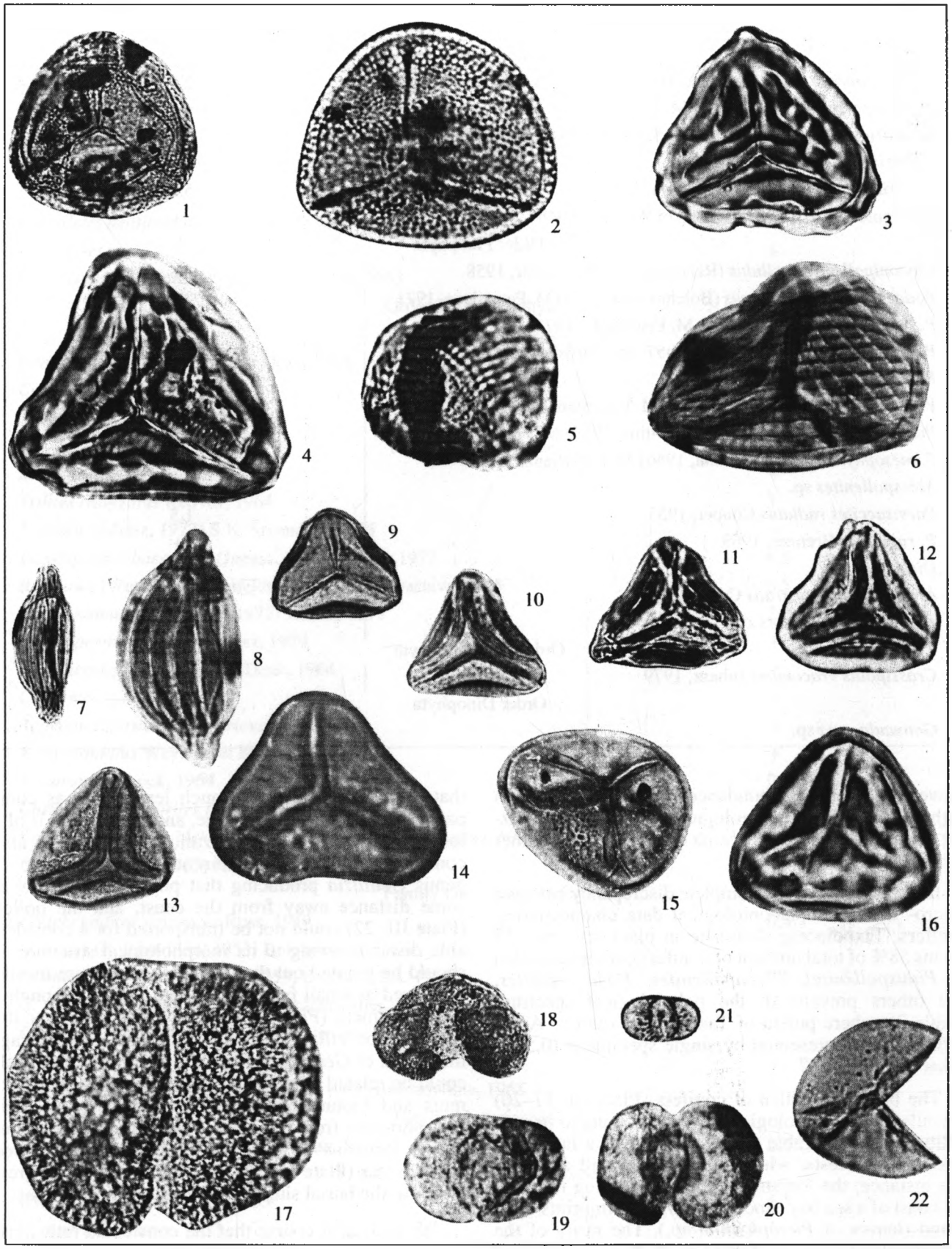


Table 3. Percentages of fossil plants and main spore and pollen groups, Exposure A_4 near the MGU training base, Prokhladnoe Village

Plant groups	Plant imprints, %	Spores and pollen, %
Lycopodiophyta	1.6	3
Equisetophyta	0.5	
Polypodiophyta	34 (27—Gleicheniaceae 3—Schizaeaceae 1—Dicksoniaceae)	51 (spores of ferns, mosses, and others) (19—Schizaeaceae 9.5—Cyatheaceae–Dicksoniaceae 4—Gleicheniaceae)
Caytoniales	2	Single grains
Pinophyta	58 (57—Taxodiaceae)	45 (44—bisaccate pollen of *Piceapollenites, Pinuspollenites, Podocarpidites*, and others) single Taxodiaceae grains
Angiospermae	4	Single grains

abundance rates of preserved palynomorphs, on the one hand, and fossil plant fragments, on the other. Nevertheless, the estimated proportions can give some idea about distribution of plant groups, and we use them.

Comparing the late Albian floras from the *Hysteroceras orbignyi* Sow. Zone of the Prokhladnoe site and from the *Pervinquieria inflata* Sow. Zone of the Kanev region (Shilkina and Doludenko, 1985; Barale and Doludenko, 1985; Doludenko and Pons, 1986; Doludenko and Teslenko, 1987; Doludenko *et al.*, 1988; Doludenko, 1992; Doludenko, 1994; Doludenko and Kostina, 1994), one can see a noticeable difference between them. The late Albian flora of the Kanev region includes two dominant families of conifers: Taxodiaceae (*Kanevia pimenova* Doludenko et Shilkina, *K. testenkoi* Doludenko, Kostina et Shilkina, *Sequoia* sp.) and Cheirolepidiaceae [*Frenelopsis kaneviensis* Barale et Doludenko, *Brachyphyllum squammosum* (Velenovsky) Palibin], which associate with rare Caytoniales (*Sagenopteris* sp.), Cycadaceae [*Dnepria schmalhausenii* (Pimenova) Doludenko et Kostina, *D. vakhrameevii* Doludenko et Kostina], and angiosperms (*Dicotylophyllum* sp.). In contrast, dominants in the flora from the Prokhladnoe region are ferns, missing from the Kanev flora, and conifers of the Taxodiaceae family only. Neither plant imprints, nor pollen of Cheirolepidiaceae were encountered in this region. (The coeval flora from the Podolia region is similar to though less diverse than that from the Kanev region.) Abundance of fossil ferns in the late Albian deposits near the Prokhladnoe Village and their absence in the Kanev region likely represent a consequence of different habitat settings. In the first case, it was presumably a sea bay, the coast of which was covered with ferns. Another story is assumed for the Kanev flora that inhabited a small island in a shallow sea, and sea transgressions probably flooded lowlands occupied by ferns approaching higher areas, from which the plant remains (conifers) were derived.

Judging from prevalence of Taxodiaceae conifers, *Gleichenia* and *Sphenopteris* ferns, the late Albian flora of the Prokhladnoe site is similar to the middle Albian flora of the Simferopol' region (Stanislavskii and Kiselevich, 1986). Moreover, the latter was found to include fern species *Phlebopteris dunkeri* (Schenk) Harris, a leaf fragment of Cycadophyta, and probably *Pterophyllum* remains.

Plate III. Spores and pollen from the lower zone of upper Albian deposits near the Prokhladnoe Village (exposure below the road bed opposite the MGU base, Sample 4857/60).

All specimens are deposited at the Paleontology Chair of the Geological Faculty, Moscow State University.

Magnification is 500 in figures 7, 8, 18, 20, and 800 in others.

(1) *Densoisporites velatus* Weyland et Krieger; (2) *Vadascisporites* sp.; (3) *Bicolisporites baconicus* (Juhasz) Juhasz; (4) *B. toratus* (Weyland et Greifeldt) S.K. Srivastava; (5) *Cicatricosisporites venustus* Deak; (6) *Cicatricosisporites* sp.; (7) *Corniculatisporites alekhinii* (Bolchovitina) Kuvaeva; (8) *C. striatus* (Deak) Kuvaeva; (9) *Gleicheniidites laetus* (Bolchovitina) Bolchovitina; (10) *G. senonicus* Ross; (11) *G. minor* Döring; (12) *G. umbonatus* (Bolchovitina) Bolchovitina; (13) *Phlebopterisporites harscutensis* Juhasz; (14) *Cyathidites australis* Coupler; (15) *Todisporites major* Couper; (16) *Duplexisporites generalis* Deak; (17) *Piceapollenites mesophyticus* (Bolchovitina) M. Petrosjanz; (18) *Pinuspollenites minimus* Kemp; (19) *Podocarpidites luteus* (Bolchovitina) M. Petrosjanz; (20) *P. multesimus* (Bolchovitina) M. Petrosjanz; (21) *Caytonipollenites* (Reissinger) Couper; (22) *Taxodiaceaepollenites* sp.

The Albian flora of Portugal (Saporta, 1894) is incomparably richer and diverse. Ferns are represented here by genera *Onychiopsis, Adiantites, Phlebopteris, Sphenopteris*, cycadophytes by genus *Pseudocycas*, conifers by *Brachyphyllum, Sphenolepis, Sequoia, Podozamites*, and angiosperms by *Aralia, Magnolia, Cissites, Menispermites, Braseniopsis, Proteophyllum, Phillites*. In addition, *Frenelopsis* remains were found in another nearby exposure of the Albian flora in Portugal.

CONCLUSION

Paleobotanical and palynological data supplementing each other elucidate composition of the late Albian vegetation. The data indicate that the coast of an ingression sea bay that corresponded to the Mangush depression was covered with diverse ferns, among which Gleicheniaceae and Schizaeaceae families dominated. Abundant shoots' fragments of genus *Geinitzia* found suggest that the coastal zone with ferns was narrow, and microphyllous forests descended close to the sea. Palynological data imply that higher slopes where occupied by conifers producing bisaccate pollen in abundance. In general, the relief was rugged and diverse, and climate was warm and humid.

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Spores, Pollen, and Organic-Walled Phytoplankton from Neogene Deposits of the Zheleznyi Rog Reference Section (Taman' Peninsula)

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Abstract—Ten spore-and-pollen assemblages and nine phytoplankton assemblages (dinocysts, green algae, acritarchs) are distinguished in upper Sarmatian–lower Kimmerian deposits of the reference section situated in the Zheleznyi Rog area of the Black Sea region. Palynological spectra of two types (forest and forest–steppe) were found to alternate throughout the section. Their composition and structure reflect climatic fluctuations, primarily the humidity changes. During humid phases, conifer (*Pinus*, Taxodiaceae) and mixed broad-leaved (*Carya, Ulmus, Quercus*) forests dominated, whereas xerophilous herbaceous–fruticose vegetation (Chenopodiaceae, Artemisia, Gramineae) prevailed during the more arid phases. A temperature decrease, most considerable in the studied stratigraphic interval, is recorded in the early Kimmerian that is evident from the increased percentage of *Sphagnum* spores and *Betula* and Ericaceae pollen in the spectra. The decrease did not exceed however thermal parameters characteristic of the warm-temperate climate and did not change it. Most significant changes in assemblages of the organic-walled phytoplankton are distinguished at two levels: in the lowermost and uppermost Pontian. The Pontian assemblage of organic-walled phytoplankton taxa *Galeacysta etrusca, "Gonyaulax digitalis," Spiniferites bentorii* and peculiar *Impagidinium*, is found to appear 8–10 m above the top of coupled diatomite layer, along which most geologists draw the boundary between the Maeotian and Pontian in the section. The assemblage is correlative with the dinocyst association from an upper part of the Messinian deposits in Italy (Lago Mare facies). In addition to taxa mentioned above, new forms, which are tentatively referred to the *Komewuia* genus, appear in the organic-walled phytoplankton assemblage of the upper Pontian and lower Kimmerian deposits. Data on salinity changes in paleobasins are discussed as well.

Key words: Black Sea region, Taman Peninsula, Zheleznyi Rog section, Sarmatian, Maeotian, Pontian, spores, pollen, dinocysts.

INTRODUCTION

Cape Zheleznyi Rog is situated on the southern (Black Sea side) coast of the Taman Peninsula. It is the site of one reference section in the northeastern Black Sea region (Fig. 1), which has been used along with others to elaborate the palyno-climatostratigraphy and regional Neogene schemes based on organic-walled phytoplankton at the Laboratory of Paleofloristics, Geological Institute RAS. The comprehensive analysis of data on palynomorphs and phytoplankton can be used to reconstruct simultaneously parameters of a basin and surrounding land, to trace dynamics of their evolution with time, to reveal synchronism and interactions of various events, and to correlate them with events in other parts of the Paratethys and Mediterranean region.

The organic-walled phytoplankton (dinocysts, green algae, acritarchs, and others) has been used in the Upper Cenozoic stratigraphy during recent decades, since the 1980s. The group is proved to be applicable for examining deposits of oceanic areas, for which zonal schemes elaborated on this basis are widely usable in practice (Williams and Bujak, 1985; Bujak and Matsuoka, 1986; Mudie *et al.*, 1990; de Verteuil, 1996; and others). In contrast to other plankton groups (nannoplankton, diatoms, and others), the organic-walled phytoplankton inhabits waters of a wide salinity range. Being independent of carbonate and silica content, it occurs in various sedimentary facies that is essential for the effective application of the group in the Upper Cenozoic stratigraphy of epicontinental, partly or completely closed basins with their specific sedimentation environments. The Paratethys (Western and Eastern) and Mediterranean Sea represented this kind of basins in various geochronological time intervals. Despite the long investigation history of studying, correlation of regional stratigraphic subdivisions in these regions is still a principal problem of their Upper Cenozoic stratigraphy.

At present, there is a great demand for elaborating zonal stratigraphic scales based on the organic-walled phytoplankton for the regions mentioned above. Work in this line is now underway, but the progress is not of equal value in different regional areas, because some of

them are under state of acquiring initial data on distribution of these microphytofossils, while in others the local stratigraphic schemes are already elaborated (Balteş, 1971; Sütő-Szentai, 1985, 1989; Ananova *et al.*, 1985; Corradini and Biffi, 1988; Al-Azhlani, 1994, and others). The organic-walled phytoplankton is most purposefully studied in the central part of Western Paratethys, especially in Hungary, where a detailed subdivision of Neogene deposits is substantiated (Sütő-Szentai, 1985, 1989) (Table 1). In Eastern Paratethys, the study of the Upper Cenozoic organic-walled phytoplankton is at the initial stage. Ananova *et al.* (1995) reported pioneering data on this group and tried to subdivide the Miocene–Pliocene deposits using data obtained in the Black Sea region, namely in the section between the Kutrya Balka and Cape Panagiya, the southern coast of Taman' Peninsula.

A keen interest to the Zheleznyi Rog section that is located on the same coast 5 km southeastward, between the Pekla Hill (Kruglaya Gora) and Cape Zheleznyi Rog, is understandable, because it is well-studied and often considered as the reference one (Andrusov, 1961 b; Popov, 1996; Popov and Zastrozhnov, 1998, and others). It is also significant that data on palynology and phytoplankton can be reliably attached here to the regional stratigraphic scheme elaborated on the basis of molluscan fauna evolution. It should be pointed out as well that Neogene deposits of the Taman' Peninsula, the Zheleznyi Rog section included, are mainly represented, in contrast to other areas of the Black Sea region, by argillaceous facies most favorable for studying both paleontological groups.

Andrusov (1961b) was the first to describe the Zheleznyi Rog section in 1899. Since that time, it attracted a considerable interest, but there is still a number of problems to be solved. The principal one concerns the boundary between Pontian and Maeotian. Andrusov pointed out that it is difficult to establish boundary between these "stages," since the Pontian *"Valenciennesia"* Clay rests on shales usually barren of molluscan remains (Andrusov, 1961a, p. 295). He described Maeotian deposits of the section (from top downwards) beginning from a thin-laminated clay member that encloses three tripoli interlayers in the upper part and a peculiar white fine-grained layer located a bit lower, resembling sand, and composed of ground volcanic glass. The member is located directly below the last clay layers, which bear small *Paradacna abichi* shells and manganese nodules and are referred to the Pontian (Andrusov, 1961a, p. 295). Thus, Andrusov places the boundary, according to fauna distribution, at the top of the mentioned member. Many of contemporari geologists prefer to place the Pontian–Maeotian boundary at the top of the first reference diatomite (tripoli) interlayers (two are well traceable along strike and third pinches out). In opinion of Yu.O. Gavrilov, it is more natural to displace the boundary 5.5–6.0 m below this level down to a thin (0.1–0.3 m) interlayer of brecciated clay ferruginate in places.

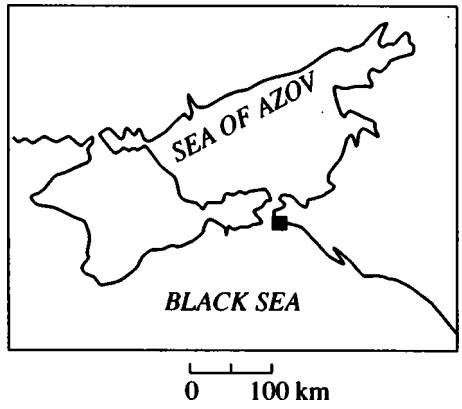


Fig. 1. Geographic location of the Zheleznyi Rog section (black box).

According to Andrusov, Pontian deposits in the Zheleznyi Rog section are about 120 m thick and can be divided in two horizons separated by a layer of *Congeria subrhomboidea* shell detritus that is 0.5–1.0 m thick; in his assessment, the upper horizon is about 40 m thick and the lower about 80 m thick (Andrusov, 1961a, p. 296; Andrusov, 1961b, p. 616). He attributed underlying deposits, without mentioning their thickness, to the Maeotian pointing to presence of *Scrobicularia tellinoides* in their lower part and emphasizing that they resemble the upper Sarmatian shales so much that they could easily be taken for the latter, because the section lacks the Maeotian demarcation horizon of bryozoan limestone (Andrusov, 1961b, p. 642). Nothing

Table 1. Subdivision of Middle–Upper Miocene deposits of Hungary on the basis of organic-walled phytoplankton (Sütő-Szentai, 1989)

Regional stage	Regional substage	Organic-walled phytoplankton zonation	
Pannonian	upper	*Mougeoita laetevirens*	
		Dinoflagellata–Zignemataceae	
		Spiniferites validus	Subzone *Spiniferites validus*
			Subzone *Spiniferites paradoxus*
	lower	*Spiniferites bentorii*	Subzone *Pontiadinium pecsvaradiensis*
			Subzone *Spiniferites bentorii oblongus*
			Subzone *Spiniferites bentorii pannonicus*
		Pleurozonaria ultima	
Sarmatian s. str.		*Spiniferites bentorii budajcnoensis*	

more detailed is reported in his book about subdivision of Maeotian deposits in the Zheleznyi Rog section.

Kimmerian deposits ("ore beds") underlying Pontian strata are described by Andrusov (from the base upward) beginning from a layer of dark brown, highly ferruginate marl (limestone in this work) that underlies a layer of bluish gray clay and next "layer of a very friable bluish gray matter" with yellow ferruginate nodules. All are overlapped by a brown iron ore bed.

FACTUAL MATERIAL

In this study, I use materials (rock samples, section description) cordially donated by S.V. Popov (Paleontological Institute, RAS), and I am very grateful to him.

According to published data (Popov, 1996; Popov and Zastrozhnov, 1998), the Zheleznyi Rog section is about 500 m thick in total. It is composed of upper Sarmatian (130 m), Maeotian (370 m), Pontian (90 m), and lower Kimmerian (10 m) deposits. A brief description of the section (Fig. 2) is presented below.

Sr_3 Member 1: gray calcareous clay alternating with yellow-gray porous limestone of vermicular texture (30 m thick).

Member 2: brown non-calcareous clay with gypsum, jarosite, and volcanic ash; the rock is exposed fragmentarily and resembles the Maikop variety; landslide marks in the lower and upper parts (100 m thick).

M_1 Member 3: gray calcareous clay with interlayers (2–3 m) of non-calcareous clay with jarosite (120 m thick).

M_2 Member 4: brecciated clay or "blocky" olistostrome (2–10 m thick).

Member 5: light gray massive calcareous clay (40 m thick).

Member 6: darker low-calcareous clay with a brownish tint (16 m thick).

Member 7: light gray thin-laminated silty high-calcareous clay, probably diatomaceous (4–5 m thick).

Member 8: light gray to greenish gray calcareous clay, brecciated at the base (23–25 m thick).

Member 9: dark gray clay with a greenish tint (5 m thick).

Member 10: light yellow thin-laminated silicified diatomite with lenses of hard limestone at the base (4.5 m thick).

Member 11: gray calcareous clay with two interlayers of diatomaceous shale and a layer (0.01 m) of gray volcanic ash 20 m above the base (40–50 m thick).

Member 12: gray to greenish gray clay with an interlayer (0.1–0.3 m) of brecciated clay at the base, in places ferruginate. Two diatomite interlayers occur 5.5–6.0 m above the base; the lower one is 0.10–0.13 m thick, and the upper is 0.03–0.06 m thick (7.0–7.5 m thick in total).

P_1 Member 13: dark gray to bluish thin-laminated calcareous clay. First *Limnocardiina* appear 3 m above the base and abundant *Paradacna abichi* occur 8 m above the base (11 m thick).

Member 14: gray massive thick-laminated clay with a reference bed of light gray diatomite (0.6 m thick) enclosing many ferruginate interlayers 2.0–2.5 m above the member base (11–12 m thick).

Member 15: dark gray banded silty clay under talus, through which some fragments are visible (13–16 m thick).

Member 16: dark gray clay with a greenish brown tint; the rock is slightly silty and calcareous (10–15 m thick).

Member 17: light to greenish gray, highly calcareous clayey siltstone (4 m thick).

Member 18: greenish gray clay with a brownish tint (2 m thick).

Member 19: Greenish lumpy to amorphous clay with whitish stains of shell detritus (1 m thick).

P_2 Member 20: greenish gray clay overfilled with shells of *Congeria novorossica, C. subrhomboidea*, and others. The contact surface with the underlying strata is wavy; the upper boundary is of erosion type, distinct owing to coloration (0.35–0.50 m thick).

Member 21: shell detritus varying in coloration from yellowish to light gray (0.15–0.23 m thick).

Member 22: greenish gray to gray clay of banded structure; bands (1–2 cm) are ferruginate, and the lower boundary is uneven (2.4 m thick).

Member 23: dark gray banded calcareous clay with brownish interlayers 0.1–0.5 m thick (9–11 m thick in total).

Member 24: lighter, brownish yellow to gray banded calcareous clay (8 m thick).

Member 25: dark gray compact calcareous clay with ferruginate brick-red interlayers responsible for banded structure (14–15 m thick).

Km_1 Member 26: dark gray limestone with a brownish tint, hard, sideritized, having uneven lower boundary (0.2–0.3 m thick).

Member 27: variegated (tobacco-green to gray) gypsum-bearing clay with ferruginate stains (0.6–0.8 m thick).

Member 28: Oolitic iron ore (3–4 m thick).

Member 29: gray, poorly calcareous clay (7–10 m thick).

It should be pointed out that due to irregular sampling some intervals of the Zheleznyi Rog section are characterized not as detailed as it is desirable. Nevertheless, the results obtained elucidate taxonomic composition of palynological and phytoplankton assemblages in separate stratigraphic units and in the whole range of studied deposits. They reveal as well the dynamic changes and main boundaries of phytoplankton renewals.

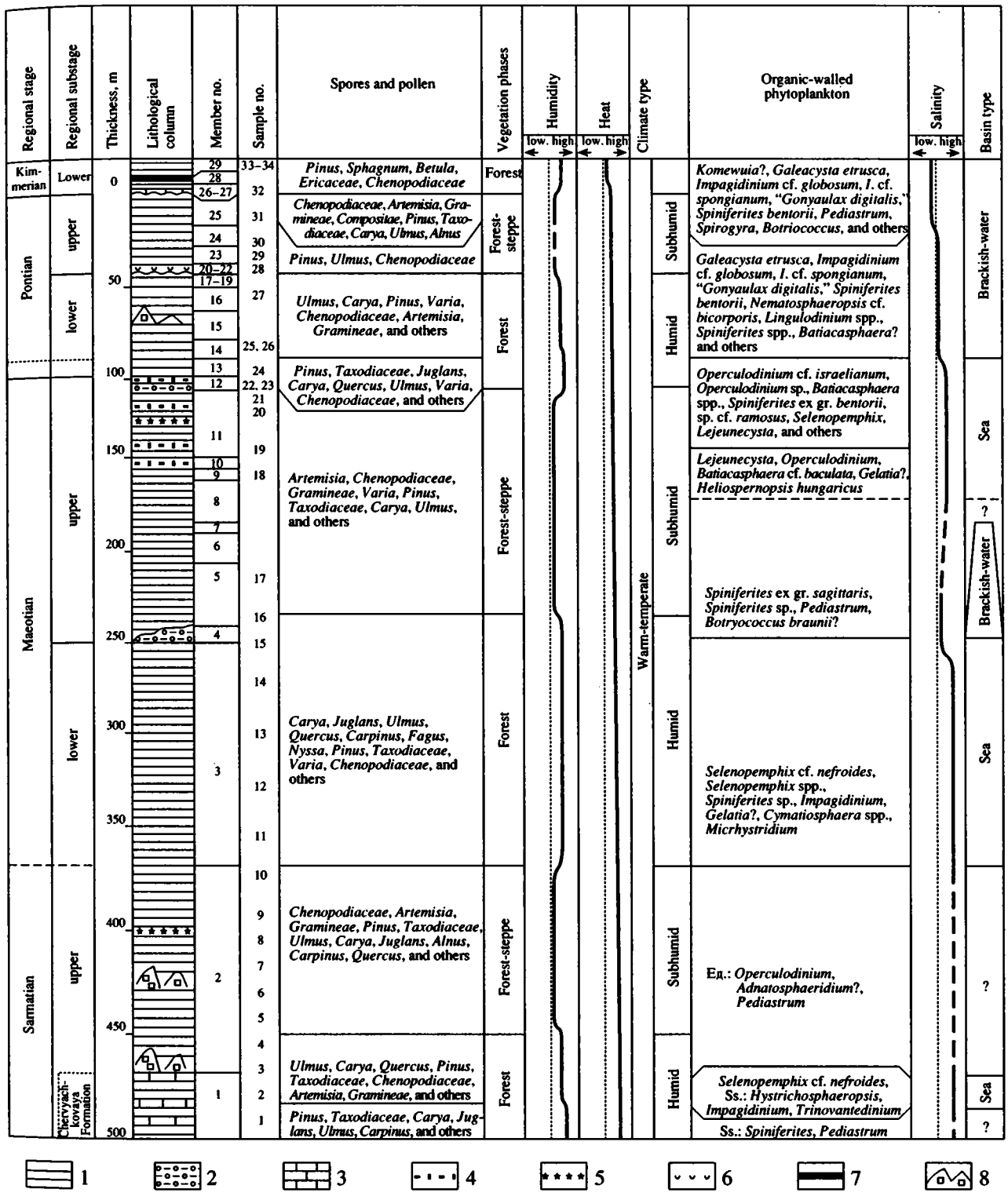


Fig. 2. Lithology and biostratigraphy of Neogene deposits in the Zheleznyi Rog section: (1) clay; (2) brecciated clay; (3) limestone; (4) diatomite; (5) volcanic ash; (6) shell detritus; (7) oolitic iron ore; (8) talus and landslides with fragmentary exposed bedrocks. Abbreviations: (ss) single specimens.

During the Seminar on Reference Neogene Sections of Eastern Paratethys that was held in August, 1998, in the Taman' Peninsula, and in field seasons of 2000–2001, the Zheleznyi Rog section was additionally sampled in more detail by a group of geologists, I belonged to. These samples will be studied in the immediate future in order to complement and verify the data available.

SPORES AND POLLEN

Ten intervals differing in palynological characteristics are distinguished in upper Sarmatian–lower Kimmerian deposits of the Zheleznyi Rog section.

The palynological assemblage from the lowermost part of the section (20 m thick, Sample 1) mostly consists of arboreal pollen; dominant here are gymnosperms: Pinaceae (*Pinus* subgen. *Diploxylon*, *Pinus* subgen. *Haploxylon*, *Abies*, *Picea*, single *Cedrus* grains), and Taxodiaceae.[1] Pollen of Juglandaceae (*Carya*, *Juglans*, *Pterocarya*), Ulmaceae (*Ulmus*, *Zelkova*), and Betulaceae (*Carpinus*, *Corylus*, *Alnus*, *Betula*) dominate among that of deciduous taxa, whereas pollen of Moraceae, *Liquidambar*, *Nyssa*, *Quercus*, *Acer*, *Salix*, *Tilia*, and Oleacea is rare. Infrequent pollen of herbaceous plants represents mainly Chenopodiaceae and less abundant *Ephedra*, *Artemisia*, Asteraceae, Cichoraceae, Gramineae, Polygonaceae, Cruciferae, Umbelliferae, Ranunculaceae and others. Single spores of *Sphagnum*, *Polypodium*, and *Lycopodium* cf. *clavatum* are also present.

Climate during the corresponding interval of the Sarmatian was warm-temperate and humid. Pine–Taxodiaceae–elm–hickory forests were widespread.

The dominating role of arboreal pollen is also characteristic of overlying upper Sarmatian deposits about 40 m thick (samples 2–4), where structure of the pollen spectrum slightly changes however. Pollen of coniferous and deciduous species is of equal abundance here, and role of elm and oak pollen increases. Chenopodiaceae still dominate among pollen of herbaceous plants, but abundance of *Artemisia* and Gramineae increases in this group. Spores are rare, representing Polypodiaceae and *Sphagnum*; individual grains belong to *Ophioglossum*, *Osmunda*, *Gleichenia*, *Lycopodium*, and *Pteridium*. Accordingly, climate was less humid as compared to the previous time interval.

In the overlying upper Sarmatian sequence about 70 m thick, the percentage of herbaceous and fruticose pollen sharply increases up more than 50% of the spore-and-pollen assemblage. In this group, Chenopodiaceae are dominants, *Artemisia* pollen is common, and abundance of Gramineae slightly decreases. Gymnosperms (*Pinus* subgen. *Diploxylon*, Taxodiaceae, *Abies*, *Picea*, *Podocarpus*, *Tsuga*) prevail among arboreal taxa. Ulmaceae and Juglandaceae (*Carya*, *Juglans*, *Pterocarya*) are dominant among deciduous forms and associate with less abundant Betulaceae (*Alnus*, *Carpinus*, *Betula*) and Fagaceae (*Quercus*, *Fagus*). Spores of Polypodiaceae, *Sphagnum*, and *Gleichenia* are rare.

At that time, climatic conditions markedly changed towards the higher aridity, and open landscapes occupied the extended areas.

Pollen of arboreal plants again plays a leading role in the lower part of Maeotian deposits (about 120 m thick; samples 11–16), where representatives of broad-leaved Juglandaceae (*Carya*, *Juglans*, single grains of *Pterocaria* and *Engelhardtia*) are dominant, elm pollen is somewhat less abundant, pollen grains of oak, hornbeam and beech are common, while *Nyssa* and other forms are frequent. *Pinus* subgen. *Diploxylon* prevailing among gymnosperms coexists with rather abundant Taxodiaceae, *Abies*, *Picea*, and with rare grains of *Podocarpus*, *Cedrus*, and *Tsuga* genera. Herbage elements (*Varia*) dominate in the herbaceous–fruticose group, where Chenopodiaceae are less abundant. Spores of Polypodiaceae and *Sphagnum* are rare. Consequently, climate was more humid during this time interval. Broad-leaved forests of elm–hickory type were widespread, and herbaceous–fruticose community was of a more mesophilic nature.

The upper Maeotian deposits (about 150 m thick; samples 17–23) are characterized by a palynological assemblage with prevailing pollen of herbaceous–fruticose vegetation. *Artemisia* and Chenopodiaceae were leading elements of the latter and associated with less abundant grasses and herbage. *Pinus* subgen. *Diploxylon* prevails in the arboreal group, where Taxodiaceae, *Carya*, and *Ulmus* are fairly abundant, commonly associated with *Quercus*, *Fagus*, *Juglans*, *Carpinus*, *Alnus*, *Picea*, *Tsuga*, and others. Spores of Polypodiaceae and *Sphagnum* are rare. Climate aridity grew in the late Maeotian, and widened areas of open landscape were occupied by steppe plant groupings, which included wormwood, Chenopodiaceae, grasses, and other elements.

The arboreal pollen with prevailing gymnosperm taxa (*Pinus* subgen. *Diploxylon*, Taxodiaceae, *Picea*, *Abies*, *Pinus* subgen. *Haploxylon*, *Sciadopitys*, *Tsuga*, *Cedrus*, *Podocarpus*, and others) again becomes dominating in spectra of lower Pontian deposits (about 10 m thick; samples 23, 24). At this level, Juglandaceae (*Juglans*, *Carya*, *Pterocarya*), oak, and elm prevail among deciduous plants, whereas alder, hornbeam, filbert, *Zelkova*, maple, holly, sweet gum, tupelo, linden, and others are less abundant. Herbaceous and fruticose pollen is represents herbage, Chenopodiaceae, *Artemisia*, grasses, and others. Amount of *Sphagnum* spores sharply increases against the background of common Polypodiaceae and *Lycopodium*. Climate of this phase was more humid than before. Pine and mixed broad-leaved forests, which included hazel nut, hickory, oak, elm, and others, dominated in the vegetation cover; swamp cypress forests grew in places.

[1] Underlined are prevailing taxa.

The palynological assemblage with prevailing arboreal pollen is also characteristic of the overlying lower Pontian deposits (about 50 m thick; samples 25–27). Leading elements of this assemblage are broad-leaved taxa *Ulmus*, Juglandaceae (*Carya, Pterocarya*), and *Quercus*. *Pinus* subgen. *Diploxylon* dominates among coniferous elements and associates with Taxodiaceae and *Abies*. The herbaceous–fruticose group includes herbage, Chenopodiaceae, wormwood, and grasses. Rare spores belong to *Sphagnum* and Polypodiaceae. As compared to the early Pontian, climate was less humid. Broad-leaved oak–hickory–elm forests were widespread.

Deposits of the uppermost Pontian, (about 25 m thick; samples 28–30), except for the top bed, are characterized by a low percentage of spores and pollen. *Pinus* subgen. *Diploxylon* is dominant taxon, whereas *Ulmus* and Chenopodiaceae are less abundant, and Taxodiaceae, *Abies, Tsuga, Picea, Carya, Pterocarya, Juglans, Quercus, Artemisia*, and others are accessory. The assemblage likely indicates the increasing aridity of climate in the second half of the Pontian.

The percentage of spores and pollen slightly increases in the topmost Pontian bed (10–15 m thick; Sample 31). Arboreal and herbaceous–fruticose pollen taxa occur here in equal proportions. The herbaceous-fruticose pollen represents Chenopodiaceae, *Artemisia*, Gramineae, Compositae, and others. *Pinus* subgen. *Diploxylon* dominates among arboreal taxa and coexists with less abundant Taxodiaceae, *Carya, Ulmus, Alnus*, and with common *Pinus* subgen. *Haploxylon, Abies, Tsuga, Picea, Pterocarya, Juglans, Quercus, Salix, Ilex, Carpinus, Betula*, and others. Polypodiaceae are main sporogenous plants, whereas spores of *Sphagnum, Pteris*, and *Gleichenia* are found as single specimens.

The prevalence of pollen *Pinus* subgen. *Diploxylon* is characteristic of the Kimmerian palynological spectra (samples 32–34). It is also remarkable that the percentage of *Betula* pollen increases at this level, whereas that of Taxodiaceae decreases. Grains of other taxa are rare, representing *Picea, Abies, Carya, Juglans, Quercus, Fagus, Alnus, Ulmus*, and others. Chenopodiaceae dominate in the herbaceous–fruticose group, where *Artemisia* and Gramineae are less abundant. The Ericaceae pollen is frequently occurring. Percentage of *Sphagnum* spores sharply increases in this interval against the background of *Lycopodium*, Polypodiaceae, and *Pteris* spores commonly present.

The increased percentage of *Betula* pollen and *Sphagnum* spores, along with the sharp decrease of Taxodiaceae and broad-leaved species, and presence of Ericaceae pollen may be considered as indications of certain cooling and relatively high humidity, although this assumption must be supported by additional data.

ORGANIC-WALLED PHYTOPLANKTON

Phytoplankton has not been found in the upper Sarmatian interval (about 130 m thick; samples 1, 3–10) of the Zheleznyi Rog section. Only single dinocysts *Spiniferites, Operculodinium, Adnatosphaeridium?* and green algae *Pediastrum* were encountered. A level with *Selenopemphix* cf. *nephroides* associated with single specimens of *Hystrichosphaeropsis, Impagidinium, Trinovantedinium*, and infrequent *Pediastrum* is distinguished in the uppermost parts of the "Chervyachkovaya Formation" (about 10 m thick; Sample 2).

As compared to upper Sarmatian deposits, Maeotian beds are richer in phytoplankton. Proximate cysts *Selenopemphix* cf. *nephroides* and *Selenopemphix* spp. dominate in the lower Maeotian interval (about 120 m thick; samples 11–15), where proximochorate cysts, e.g., smaller stunted *Spiniferites* sp. and *Impagidinium?* are also common. Green algae *Gelatia?, Cymatiosphaera* spp. are abundant here, and acritarchs *Micrhystridium* sp. are also present.

The lower part of the upper Maeotian deposits (about 80 m thick; samples 16, 17) yields the phytoplankton assemblages of abundant small stunted *Spiniferites* ex. gr. *sagittarius* with reduced spines, *Spiniferites* spp., and rare *Impagidinium*. Other forms frequently occurring here are *Pediastrum* and *Botryococcus braunii?*.

Another assemblage from the overlying interval (30–35 m; samples 18, 19) consists of rare *Lejeunecysta* spp. and single representatives of *Operculodinium* sp. and *Batiacasphaera* cf. *baculata*. Green algae *Gelatia?* are abundant here, and *Heliospermopsis hungaricus* is encountered.

The upper Maeotian–lower Pontian boundary interval (40 m thick in total; samples 20–24) is characterized by the assemblage, in which larger *Operculodinium* cf. *israelianum* dominate and associate with less abundant *Operculodinium* sp., *Batiacasphaera micropapillata?, B. sphaerica?*, and *B. baculata?*. Dwarfed *Spiniferites* cf. *bentorii, S.* cf. *ramosus, S.* cf. *cingulatus, Spiniferites* spp., *Cleistosphaeridium?, Lingulodinium, Selenopemphix* sp., *Lejeunecysta* sp., *Impagidinium*, and *Achomosphaera* ex gr. *sagenata?* are common in this interval. Other taxa abundant here are green algae *Gelatia?* and *Cymatiosphaera. Pediastrum, Botryococcus, Micrhystridium* sp. are less abundant, while *Pterospermopsis, Heliospermopsis*, and *Spirogira* with ornamented surface are found as single specimens.

The phytoplankton assemblage from overlying layers of lower Pontian deposits (Member 14) is remarkably different. It marks the appearance level of *Galeacysta etrusca* and of a new group of peculiarly sculptured *Impagidinium* forms (*I.* cf. *spongianum, I.* cf. *globosum, Impagidinium* sp.), which have reduced sutural crests (septae). The last group dominates in phytoplankton assemblages of the whole Pontian interval (about 65 m thick; samples 25–30), except for its lowermost (8–10 m thick) and uppermost parts. Associated

Plate 1. Organic-walled phytoplankton from Neogene deposits of the Zheleznyi Rog section (magnification ×750 for figures 1–3, 5–10, and ×500 for figure 4)

(1, 2) *Impagidinium* cf. *globosum* Sütő-Szentai, 1985, Sample 25, Pontian; (3) *Impagidinium* cf. *spongianum* Sütö-Szentai, 1985, Sample 25, Pontian; (4) *Gonyaulax digitalis* (Pouchet, 1883) Kofoid, 1911 (see Footnote 2), Sample 25, Pontian; (5) *Spiniferites bentorii* (Rossignol, 1964) Wall and Dale, 1970, Sample 29, Pontian; (6) *Spiniferites* cf. *ramosus* (Ehrenberg, 1838) Mantell, 1854, Sample 30, Pontian; (7) *Spiniferites* sp., Sample 15, Maeotian; (8) *Spiniferites* sp., Sample 30, Pontian; (9) *Galeacysta etrusca* Corradini and Biffi, 1988, Sample 32, Kimmerian; (10) *Galeacysta etrusca* Corradini and Biffi, 1988, Sample 25, Pontian.

taxa are *"Gonyaulax digitalis,"*[2] *Spiniferites bentorii, Nematosphaeropsis* (including *N.* cf. *bicorporis, Nematosphaeropsis* sp.). *Lingulodinium* spp., various *Spiniferites* forms (*S.* ex gr. *cornutus, S.* ex gr. *elongatus, S.* ex gr. *perforata, S.* ex gr. *ramosus, S.* aff. *frigidus*), and *Batiacasphaera?* species (*B. sphaerica?, B. micropapillata?, B. baculata?*). *Selenopemphix, Adnatosphaeridium* cf. *multispinosum, Hystrichokolpoma* spp., *Impagidinium* cf. *pallidum, Cannosphaeropsis?, Cerebrocysta?, Achomosphaera, Apteodinium, Hystrichosphaeridium, Hystrichosphaeropsis* sp., *Operculodinium* are present as single specimens. *Pediastrum* occurs frequently in contrast to less common smooth *Spirogyra.*

In the upper Pontian deposits (about 15 m thick; Sample 31), the phytoplankton assemblage again substantially changes in composition. New taxa appearing here are similar to *"Diconodinium inequicornutum"*[3] from Pontian deposits of Hungary (Baltes, 1971) and to *"Pontiadinium"*[4] forms from the mid-Pannonian boundary deposits of Hungary (Sütö-Szentai, 1985). This group of taxa is characteristic, along with *Galeacysta etrusca* and peculiar *Impagidinium* forms, of the overlying Kimmerian deposits (about 10 m thick; samples 32–34). The assemblage of the late Pontian–early Kimmerian time also includes *"Gonyaulax digitalis," Spiniferites bentorii, Nematosphaeropsis* sp., *Spiniferites* ex gr. *cornutus, S. ramosus, Spiniferites* spp., *Hystrichosphaeropsis, Batiacasphaera?, Lingulodinium, Hystrichosphaeridium, Achomosphaera?, Leptodinium?, Apteodinium.* Representatives of *Pediastrum, Spirogyra* (ornamented and smooth morphotypes of the latter), and *Botryococcus.*

DISCUSSION

As a result of palynological study, ten palynological assemblages have been established in upper Sarmatian–lower Kimmerian deposits of the Zheleznyi Rog section.

Three palynological assemblages are distinguished in upper Sarmatian deposits. Two of them exemplify development of mesophilous forest vegetation in areas adjacent to the Euxinian Basin, whereas the third one corresponds to xerophilous forest–steppe plant communities. At the beginning of the late Sarmatian (Sample 1) climate was warm-temperate and humid. Forests with Pinaceae, Taxodiaceae, Ulmaceae, and *Carya* were widespread at that time. In the subsequent epoch of the late Sarmatian (samples 2–4), climate was less humid. Substantial changes occurred at the end of the Sarmatian (samples 5–10) when the climate became drier, and the open landscapes areas considerably extended.

Maeotian deposits are characterized by two palynological assemblages: the older one corresponding to a forest phase in vegetation evolution and the younger indicative of the forest–steppe phase. The climate humidity increased in the first half of the Maeotian (samples 11–16) as compared to the latest Sarmatian time. Broad-leaved elm-hickory forests were widespread, and herbaceous–fruticose plants were of a more mesophilous type. In the late Maeotian (samples 17–22), when climatic conditions were more arid, the open landscape areas extended and hosted steppe-type plant groupings with *Artemisia*, Chenopodiaceae, and others grasses.

Four palynological assemblages characterize Pontian deposits. The first and second assemblages of the early Pontian time point to the prevalence of mesophilous forest formations in the vegetation cover, whereas the other two, corresponding to the late Pontian, characterize a more xerophilous forest–steppe plant communities. At the beginning of the Pontian (samples 23, 24), the climate humidity increased as compared to the late Maeotian time. Pine and mixed broad-leaved forests with hazel nut, hickory, oak, and elm dominated in the vegetation cover, but in places there were forests with *Taxodium.* The later interval of the early Pontian (samples 25–27) corresponds to a slightly less humid climate as compared to the initial Pontian time. Broad-leaved forests with oak, hickory, and elm were widespread at that time. It is likely that climate was drier in the second half of the Pontian (samples 28–30), and herbaceous–fruticose formations occupied larger areas.

One palynological assemblage of the early Kimmerian time indicates prevalence of arboreal vegetation with dominating pine formations. Judging from the leading role of sphagnum moss, heath, and birch, whereas abundance of *Taxodium* and broad-leaved spe-

[2] Name *Gonyaulax digitalis* is used to denote a recent dinoflagellate species in its mobile stage, and it is not legitimate to apply this name for morphologically similar fossil cysts, that is why it is used with quotation marks. The cyst species under consideration likely bear no relation to *Gonyaulax digitalis* at all, and *Spiniferites bentorii* can be a fossil analog of the latter (Dodge, 1989).

[3] At present, this species is attributed to the genus *Komewuia* (Chen, 1982; Williams *et al.*, 1998).

[4] *"Pontiadinium"* is recognized as a later synonym of *Komewuia* (Chen, 1982; Williams *et al.*, 1998).

Plate I

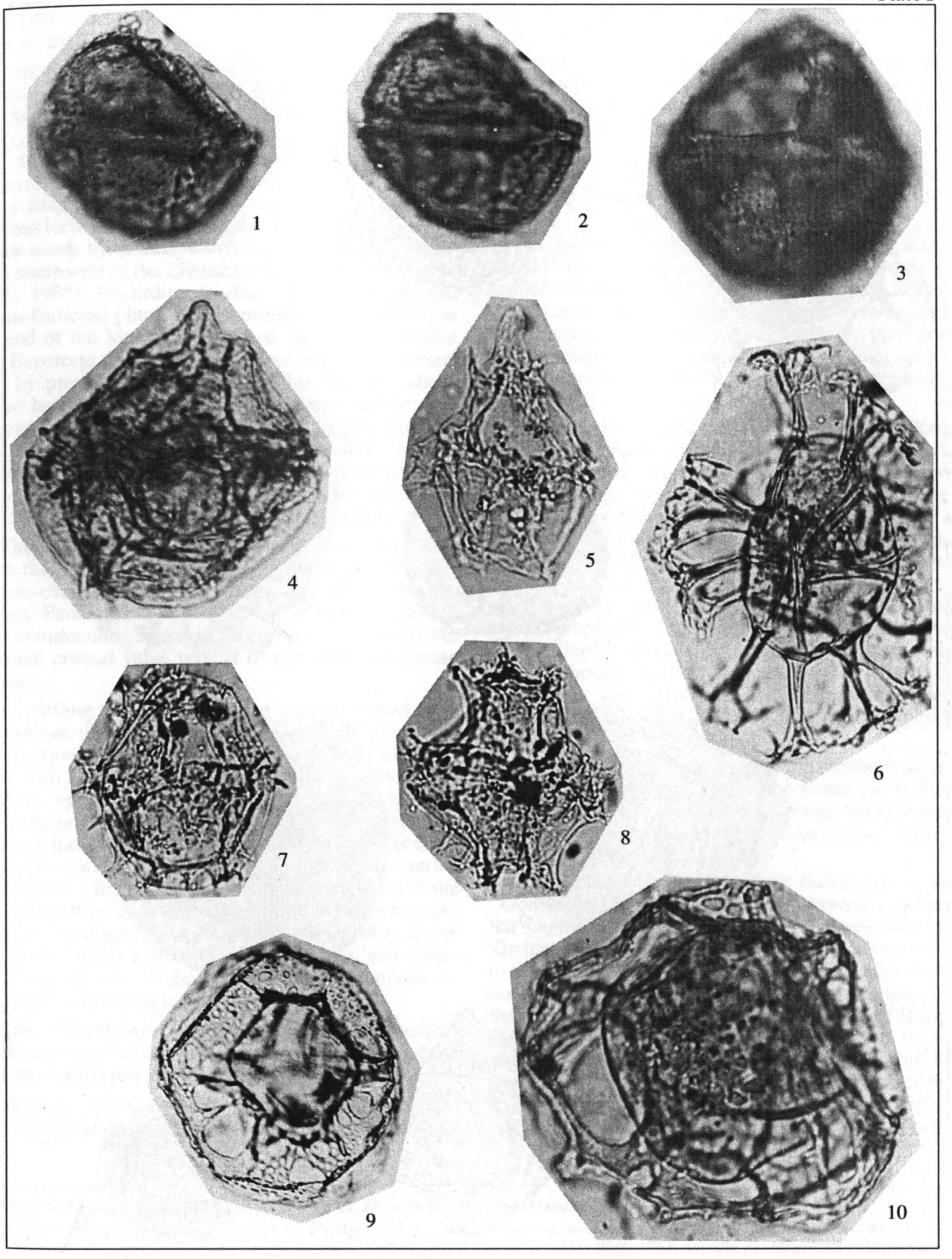
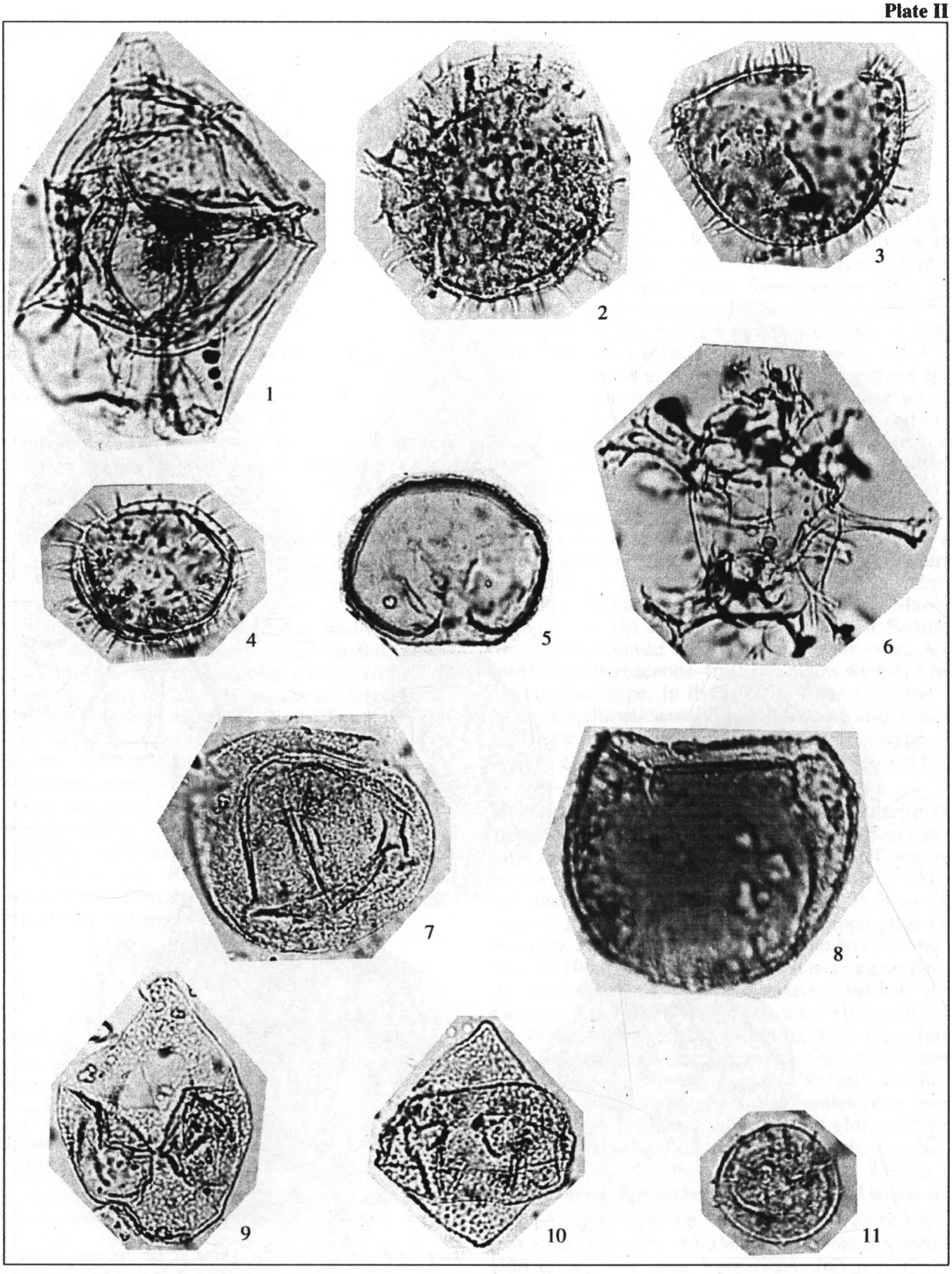


Plate II



cies decreased, the early Kimmerian climate was likely more cold and humid, although this assumption needs to be verified.

Hence, this study revealed alternation of forest and forest–steppe palynological assemblages within the late Sarmatian–early Kimmerian interval and, respectively, the regular change with time of humid and more arid climatic phases. The data obtained are in a good agreement with the results of palynological studies of upper Maeotian, Pontian, and Kimmerian deposits in sections located between the Cape Panagiya and Kutrya Balka along the southern coast of the Taman' Peninsula, southwest of the Zheleznyi Rog section (Ananova *et al.*, 1985). According to data by Ananova, herbaceous–fruticose plants also dominated in vegetation at the end of the Maeotian. The first half of the Pontian (the Evpatoria, Ingulets, and Odessa beds) is characterized by prevalence of arboreal plants, whereas abundance herbage and fruticose taxa again significantly increases in the second half of this stage (the Portafer and Bosporus beds). The early Kimmerian spectra (those from the Azov Horizon) show a higher percentage of pollen produced by small-leaved forests (birch, alder, willow) that indicates the climate cooling.

Thus, data by Ananova and results presented above show that most substantial cooling during the late Sarmatian–early Kimmerian period occurred in the Taman' Peninsula at the beginning of the Kimmerian. This temperature decrease likely did not exceed the minimal critical value typical of the warm–temperate climate.

According to new data on the organic-walled phytoplankton, the upper Sarmatian part of the Zheleznyi Rog section is divided in two intervals by a thin interlayer with *Selenopemphix* cf. *nefroides* (Sample 2). At present, species *Selenopemphix nefroides* inhabits inner regions of the neritic zone within temperate latitudes of the Atlantic Ocean (Harland, 1983). Accordingly, we may suggest that there was a short marine phase in the late Sarmatian, which corresponded to the deposition time of uppermost beds of the "Chervyachkovaya Formation." Phytoplankton was found as single specimens below and above this interlayer, and corresponding periods of the upper Sarmatian apparently were unfavorable for its development.

The overlying Maeotian–lower Kimmerian sequence is characterized by six diverse assemblages of organic-walled phytoplankton.

Various *Selenopemphix* species, indicating that salinity of paleobasin was close to normal, dominate in the lower Maeotian assemblage (samples 11–15). Dwarfed *Spiniferites* species and frequent occurrence of fresh-water phytoplankton (*Pediastrum, Botryococcus braunii?*) in the assemblage from the lowermost part of upper Maeotian (samples 16, 17) are indicative of a certain freshening at the beginning of the late Maeotian. The taxonomic composition of the assemblage changes across the mid-late Maeotian boundary (samples 18, 19), above which marine forms *Lejeunecysta, Batiacasphaera* cf. *baculata*, and others appear.

The end of the late Maeotian (samples 20, 21) and initial Pontian (samples 22–24) are characterized by the assemblage mainly consisting of marine and brackish-water forms, such as *Operculodinium* cf. *israelianum, Selenopemphix, Lejeunecysta, Spiniferites*, taxa similar to *Batiacasphaera* species, and others, which associate with few freshwater forms (*Pediastrum, Botryococcus*, and *Spirogira* with ornamented surface).

It should be pointed out that the Maeotian–Pontian boundary, if one assumes to trace it in the coupled diatomite layer, would be inside the range of this indivisible assemblage. Significant changes in composition of the organic-walled phytoplankton, i.e., appearance and bloom of many taxa (*Galeacysta etrusca, "Gonyaulax digitalis," Spiniferites bentorii*, peculiar *Impagidinium* species, and *Nematosphaeropsis* cf. *bicorporis*), took place 8–10 m above the top of the coupled diatomite layer (Sample 25). The Pontian molluscan fauna (*Paradacna abichi*) appears in mass roughly at the same level, although its first specimens were found 5 m below it as well. Mollusks and phytoplankton populated a brackish-water basin during the corresponding period.

In the section exposed along the southern coast of the Taman' Peninsula, from Cape Panagiya to Kutrya Balka, Ananova *et al.* (1985) correlated the appearance level of the assemblage with "*Gonyaulax*" with the lower part of the Odessa Beds.

The dinocyst assemblage with *Galeacysta etrusca, "Gonyaulax digitalis," Spiniferites bentorii*, and peculiar *Impagidinium* forms, which is distinguished in the Zheleznyi Rog section (samples 25–30), is very similar to that of the upper *Pontiadinium pecsvaradiensis* Subzone of the *Spiniferites bentorii* Zone from the uppermost part of the lower Pannonian deposits of Hungary (Sütő-Szentai, 1985), where it corresponds to acme zones of *Spiniferites bentorii, "Gonyaulax digitalis,"* and *Pontiadinium pecsvaradiensis*. On the other hand, it is analogous to the assemblage of the *Spiniferites validus* Zone in upper Pannonian deposits ("Postpan-

Plate II. Organic-walled phytoplankton from the Zheleznyi Rog section (magnification ×750 for all figures).

(1) *Hystrichosphaeropsis* sp., Sample 30, Pontian; (2) *Operculodinium* cf. *israelianum* (Rossignol, 1962) Wall, 1967, Sample 21, Maeotian; (3) *Operculodinium* cf. *centrocarpum* (Deflandre and Cookson, 1955) Wall, 1967, Sample 21, Maeotian; (4) *Operculodinium* sp., Sample 21, Maeotian; (5) *Selenopemphix* sp., Sample 21, Maeotian; (6) *Dinoflagellate* gen. indet., Sample 30, Pontian; (7) *Dinoflagellate* gen. indet., Sample 20, Maeotian; (8) *Palaeostonocystis*?, Sample 28, Pontian; (9) *Komewuia*?, Sample 31, Pontian; (10) *Komewuia*?, Sample 30, Pontian; (11) *Cymatiosphaera* sp., Sample 15, Maeotian.

Table 2. Correlation of late Miocene–early Pliocene organic-walled phytoplankton assemblages from the Black Sea region, Western Paratethys, and Mediterranean region

Eastern Paratethys		Western Paratethys				Mediterranean Sea	
Black Sea Basin (after Filippova)		Hungary (after Sütő-Szentai, 1985, 1989)		Romania (after Baltes, 1971)		Italy (after Bertini and Corradini, 1998)	
Lower Kimmerian	*Komewuia*?, *Galeacysta etrusca*, *Impagidinium* cf. *globosum*, *I.* cf. *spongianum*, and others		Assemblage of the *Mougeotia laetevirens* Zone	Pontian	*Diconodinium inequicornutum** and others		
			Assemblage of the *Dinoflagellata–Zygnemataceae* Zone				
Pontian	*Galeacysta etrusca*, *Impagidinium* cf. *globosum*, I. cf. *spongianum*, and others	Upper Pannonian (Pontian)	Assemblage of the *Spiniferites validus* Zone (*Impagidinium globosum*, *I. spongianum*, *S. paradoxus*, and others)			upper Messinian (Lago Mare)	*Galeacysta etrusca*, *Impagidinium* sp. ("pitted"), and others

* See Footnote 3 in the text.

nonische neogen Sedimente," "Oberpannonische Unterstufe," "pontischen Bildungen") of Hungary (Sütő-Szentai, 1985, 1989), where this zone is distinguished at the appearance level of *Spiniferites validus*, corresponds to *Impagidinium globosum* and *I. spongianum* acme zones, and also yields "*Gonyaulax digitalis*," *Spiniferites bentorii*, "*Pontiadinium*" species, and other forms.

At this stage in my research, I tend to correlate the indicated assemblage from the Zheleznyi Rog section with the upper Pannonian assemblage of the *Spiniferites validus* Zone, taking into account abundance of *Impagidinium* cf. *globosum* and *I.* cf. *spongianum* (Table 2) at this level. This assumption needs, however, to be verified.

In the Mediterranean region (Italy), the assemblage with *Galeacysta etrusca* and "pitted" *Impagidinium* forms (Bertini and Corradini, 1998) appears in the upper part of the Messinian (Lago Mare facies) and is correlative with the Pontian dinocyst assemblage of the Black Sea region.

One more appearance level of new taxa, which are similar to *Komewuia* species (see Footnote 2), is found in the Zheleznyi Rog section at the top of Pontian deposits (Sample 31). They associate here with peculiar *Impagidinium, Galeacysta etrusca, "Gonyaulax digitalis,"* and *Spiniferites bentorii* inherited from the previous assemblage. The same assemblage is characteristic of lower Kimmerian layers as well. The above taxa developed most likely under brackish-water conditions. Judging from the presence of freshwater forms (*Pediastrum, Spirogira, Botryococcus*), the paleobasin became slightly freshened at the beginning of the Kimmerian that might have been related to the increase of river runoff and climatic cooling that is inferable from palynological data on that interval of the Kimmerian.

Two levels of changes in the organic-walled phytoplankton composition, which are established in the lowermost and uppermost parts of the Pontian in the Zheleznyi Rog section, are traceable as well in the section located eastward of the Kuban superdeep borehole.

CONCLUSION

Hence, ten spores-and-pollen assemblages of the forest and forest–steppe type, which reflect alternation with time of corresponding vegetation phases, are established after palynological analysis of upper Sarmatian, Maeotian, Pontian, and lower Kimmerian deposits in the Zheleznyi Rog section. In the entire studied interval, the assemblages are nearly identical in taxonomic composition (at the level of families and genera) and differ only in proportions of those components, which reflects climatic fluctuations.

At the early phases of the late Sarmatian, Maeotian, and Pontian, climate was warm-temperate and humid with coniferous (of pine and *Taxodium* types) and mixed broad-leaved forests (elm–hickory, oak–hickory–elm, and others) dominated. During later phases of the mentioned ages, climate was more arid, and areas of xerophilous herbaceous–fruticose plants, mostly of Chenopodiaceae, *Artemisia*, and other grasses, widened. Judging from a considerable increase in the abundance of such taxa as *Sphagnum*, Ericaceae, and *Betula*, climatic conditions became more cool and relatively humid at the beginning of the Kimmerian.

Palynological data reveal most clearly the rhythmic changes in the humidity. Alternating forest and forest-steppe vegetation phases corresponded to humid and subhumid conditions, respectively. Variations in temperature developed in other way, showing a general tendency of gradual cooling that is evident from a steadily

decreasing percentage of spores and pollen of thermophilic plants upward in the section. During the entire studied period, climate was likely warm-temperate. The most remarkable decrease in the temperature apparently did not exceed the minimal values typical of the warm-temperate climate.

Nine assemblages of the organic-walled phytoplankton (dinocysts, green algae, acritarchs) have been established in the course of paleoalgological study of rocks from the Zheleznyi Rog section.

The late Sarmatian is poorly characterized by phytoplankton, and only one short interval of a definite marine regime is distinguishable here. The basin salinity in the early Maeotian was close to normal and decreased at the beginning of the late Maeotian. The marine regime recommenced in the second half of the late Maeotian, but water salinity likely deviated from normal at that time. Brackish-water conditions were characteristic of the Pontian and early Kimmerian, and the lowest salinity parameters are recorded by the uppermost Pontian and lowermost Kimmerian phytoplankton assemblages. Freshening events at the early late Maeotian, terminal Pontian, and earliest Kimmerian likely reflected activation of the river runoff. The last event could be also connected with the temperature decrease inferable from palynological data.

Most radical changes in composition of phytoplankton assemblages, which are marked by appearance of many new taxa within the late Sarmatian–early Kimmerian interval, took place in the lowermost (*Galeacysta etrusca*, peculiar *Impagidinium*, and other forms) and uppermost *(Komewuia?)* Pontian. These levels are good stratigraphic markers traceable in easterly areas of the Black Sea region (Kuban superdeep borehole).

According to distribution patterns of organic-walled phytoplankton, the Maeotian–Pontian boundary in the Zheleznyi Rog section is actually traceable 8–10 m above the top of the coupled diatomite layer, to which it was conventionally attributed.

The results of correlation of the late Neogene organic-walled phytoplankton assemblages from the Black Sea region with coeval assemblages from the Mediterranean region and Western Paratethys are quite significant. They imply that the Pontian assemblage with *Galeacysta etrusca* and peculiar *Impagidinium* forms, which are characteristic of the greater interval of Pontian deposits in the Black Sea region (except for its lowermost and uppermost parts), is correlative with the late Messinian assemblage (Lago Mare facies) of the Mediterranean region and, most likely, with the assemblage of the late Pannonian (Pontian) *Spiniferites validus* Zone of Hungary. The assemblage of organic-walled phytoplankton from uppermost Pontian and lower Kimmerian deposits of the Black Sea region exhibit similarity with the Pontian assemblage of Romania.

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Accumulation Conditions and Correlation of the Upper Eocene Amber-Bearing Prussian Formation, Kaliningrad Region

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Abstract—Accumulation conditions of largest (about 90% of known world reserves) amber deposits in the upper Eocene Prussian Formation of the Sambiiskii Peninsula, Kaliningrad region, is discussed. Microfossils and findings of amber, glauconite, and phosphorite are used to correlate local stratigraphic units of the late Paleogene in the Kaliningrad region, Belarus, Ukraine, Poland, and Germany, which correspond or are close in age to the Prussian Formation stratotype situated in the Sambiiskii Peninsula.

Key words: Paleogene, Eocene, Prussian Formation, amber, amber-bearing deposits, phosphorite, glauconite, foraminifers.

INTRODUCTION

The Paleogene sandy to clayey terrigenous facies are widespread in the western Kaliningrad region of Russia and in adjacent Baltic Sea areas, namely, in the Vistula and Courland lagoons. They rest on various Upper Cretaceous horizons and are overlain by Neogene or Quaternary deposits. The upper Eocene sediments bearing amber and phosphorites, are of a high commercial importance. The Prussian Formation, first recognized in the XIX century by Zaddach (1868), is the major productive subdivision of the upper Eocene. The amber-bearing sediments of the same age are also known in Denmark, northern Germany, Poland, Belarus, and Ukraine, though they usually differ in lithology and facies affinity.

In this work, our objective was to examine data on facies environments, accumulation history, and stratigraphy of amber-bearing deposits in the Kaliningrad region and to correlate them with age-comparable stratigraphic units known in the regions mentioned above. In addition to published data, we used new data on oil-prospecting boreholes drilled in the Kaliningrad region and Baltic Sea. We also studied the Prussian Formation sections exposed in new quarries of the Amber factory, which are located in the Sambiiskii Peninsula, the western Kaliningrad region.

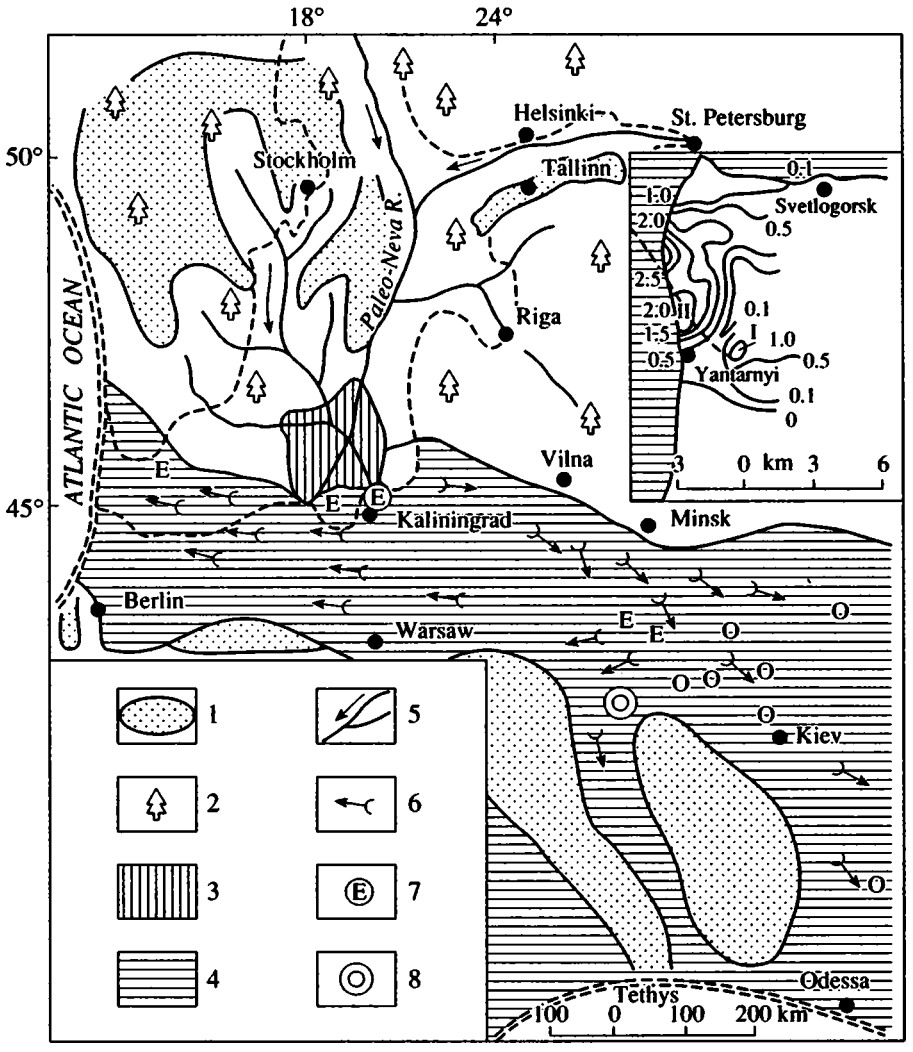
CONDITIONS AND ACCUMULATION HISTORY OF AMBER-BEARING DEPOSITS

The largest amber deposit associated with the "Blue Earth" bed of the upper Eocene Prussian Formation, is situated in the northwestern Kaliningrad region near the Yantarnyi Settlement. The amber-bearing sediments are attributed to deltaic deposits of a river that drained Scandinavia (Katinas, 1971). In the Paleogene, the greater part of Baltic region and Scandinavian Peninsula was covered by coniferous forests. Faunal remains enclosed in amber suggest a warm and humid climate of the "amber" forests.

Fossil insects are mostly represented by fungus midges (Mycetophilidae, Nematocera) whose larvae develop in fungi or rotten plant remains. Long-legged flies (Dolichopodidae, Brachycera), which can be encountered in grass or on bush leaves near the water, are also widespread. Numerous caddis flies whose larvae develop only in the water also indicate a high humidity of the "amber" forests. Findings of termites, Acrididae, and Mantoidea suggest that climate of Fennoscandia was warmer than nowadays (Plavil'shchikov, 1994). Remains of rare and specific desert or steppe insects (tetracampids and encirtids of Hymenoptera order), which are typical nowadays of arid regions and found enclosed in Baltic amber (Trjapitzin and Manukyan, 1995), are most likely a result of eolian transportation from southern regions adjacent to the Tethys area.

Strengthened monsoon winds of the late Eocene time broke trees, which oozed resin. The resin accumulated in soils was transported then by creeks and rivers to the major stream that discharged, in the northwestern Kaliningrad region, into a wide strait connecting the North Atlantic with the Tethys (see the figure).

Indications of anaerobic sedimentation environments, e.g., pyritization, smell of hydrogen sulphide, and black color, which are characteristic of amber-bearing sediments, suggest their lagoonal or lagoonal-deltaic origin (Kharin, 1995). This is also evident from data on amber distribution in the upper Eocene sediments ("Blue Earth") within the prospected area of the Sambiiskii Peninsula. The maximum amber content, >1 kg/m^3, is confined to the lagoon axial zone and



Paleogeographic scheme of the distribution area of upper Eocene and lower Oligocene amber-bearing deposits: (1) upland; (2) marshy lowland covered by coniferous forests; (3) lagoonal and deltaic facies with primary amber deposits; (4) strait connecting the Tethys with seas of the Atlantic and Boreal oceans; (5) river system that transported resin to lagoons and strait; (6) direction of amber dispersal toward accumulation sites of secondary (placer) amber deposits; (7) Eocene; and (8) lower Oligocene large amber deposits sediments (letters E and O designate small deposits). Dashed line indicates contours of the present-day Baltic Sea.

Isocons in the inset map show abundance rate of amber (kg/m^3) in the "Blue Earth" and outline the axial zone of the past lagoon in the Sambiiskii Peninsula. Roman figures indicate localities of the Primorskii (I) and Plyazhevyi (II) quarries.

increases northward of the Primorskii to Plyazhevyi quarries (see inset map in the figure). The lagoonal amber-bearing sediments, such as the black pyritized clayey siltstones intercalated with clay and sand, are distinguished in the Plyazhevyi quarry of the Amber factory. The quarry in the bottom area of the Baltic Sea fenced off by dam, is located to the north of the Yantarnyi Village. As is visible in the quarry, the lagoonal amber-bearing deposits gradually decrease in thickness toward the northwest, and, according to the available data (Blazhchishin *et al.*, 1978; Kharin *et al.*, 1991; Kharin, 1995), they completely wedge out 10–15 km away from the current shoreline at a depth of 15 to 20 m.

Geophysical profiles suggest the replacement of lagoonal amber-bearing facies the channel ones at the bottom of the sea, where ancient incisions in the Mesozoic sediments are detected (Sviridov, 1984). Dimensions of the incised valleys (up to 1 km wide and to 100 m deep), and boulders and shingle lenses intercalated with the lagoonal and lagoonal-deltaic amber-bearing deposits indicate a considerable size of the late Eocene Paleo-Neva river (Kharin, 1995). Judging from the clastic material composition, the river drained not only the Baltic Sea bottom and islands Gotland, Saaremaa, and others, but the Scandinavian shield as well. It transported clastic material and resin of coniferous trees that was still non-petrified. The resin accumulated, experienced fossilization, and turned into amber

in the lagoon and delta thus forming the primary amber deposits.

The anaerobic conditions in lagoons, where resin was transformed into amber, were rather important from the standpoint of the amber quality. The resin that escaped the ripening stage in such environment and accumulated on the sea bottom, turned into amber of inferior quality ("raw," "wet," "soapy," or "young").

During sea-level oscillations the amber deposits of delta and lagoons were eroded by alongshore currents, direction of which depended on that of monsoon winds, southeastern in winter and northwestern in summer (Cincura, 1989).

A wide dispersal of amber resulted from its low density and high hydraulic size. Since amber was transported at a long distance from primary resin burials and formed secondary accumulations, it may be considered as a placer mineral (Trofimov, 1974). The amber dispersal within the Eocene and early Oligocene sea resulted in formation of secondary amber deposits in Poland, Germany, Denmark, Belarus, and Ukraine, occurring in the latter as far to the south as the Carpathians (Katinas, 1971). Amber from these countries is close in characteristics and composition to that from the Sambiiskii Peninsula. When the Mazury–Belarusian uplift dammed the strait transforming it into a bay, amber accumulated at the northern slope of the uplift, and the large amber placer deposits may be expected here. Together with amber, the currents transported different minerals, glauconite included. The phosphorite-bearing bed at the base of amber-bearing sediments of the Prussian Formation was likely formed in response to water exchange between the Tethys and North Atlantic, which led to penetration of deep phosphate-rich waters into the strait (Kharin, 1995, 1997).

Outside the Sambiiskii Peninsula, the amber-bearing deposits are of diverse facies affinities. In the Chlapowo deposits of northeastern Poland, their lithologic composition is close to that of the Sambiiskii region. Accordingly, it was inferred that the Sambiiskii and Chlapowo (I–II) amber-bearing sequences originated in the same delta (Kosmowska-Ceranowich and Muller, 1985). However, the detailed prospecting showed that the productive "Blue Earth" does not occur to the west and southwest of the Sambiiskii Peninsula. Most likely, there were different marine lagoons of the same age, the site of the Paleo-Neva branches discharge. In Ukraine and Belarus, amber fragments are characteristic of the terrigenous glauconitic sediments (Katinas, 1987), but they were also found in the carbonate and clayey-carbonate (marly) deposits.

STRATIGRAPHIC POSITION AND CORRELATION OF THE PRUSSIAN FORMATION

The Prussian Formation is a local stratigraphic unit distinguished by Zaddach according to several indications, the major among which are lithologic composition, faunal and floral remains, and occurrence of amber, glauconite, and phosphorite concretions. In the stratotype at the Sambiiskii Peninsula, the formation is up to 50 m thick. The following four beds are distinguished here (from the base upward).

(1) "Wild Earth:" clayey, inequigranular, glauconite–quartz sand bearing clay segregations, abundant phosphorite concretions, and shark teeth redeposited from the underlying sediments of the Alk Formation.

(2) "Blue Earth:" greenish gray to greenish black, clayey, micaceous, pyritized, glauconite–quartz sand and silt with abundant amber fragments, phosphorite and pyrite concretions, wood remains, and pyritized fucoids, the traces of worm life activity.

(3) "Quicksand:" greenish-gray inequigranular sand bearing siderite and pyrite nodules, some enclosing amber; phosphorite concretions and clayey lumps occur as well. In the eastern Sambiiskii Peninsula (the Svetlogorsk–Primor'e region), the lower part of the bed is represented by ferruginous sandstones known under the name "Krant."

(4) "White Wall:" highly micaceous silt with carbonized plant remains and small amber fragments.

The most common criterion used in the amber prospecting, the presence of "Blue Earth," is not a universal one, because amber was dispersed by currents in the Eocene and early Oligocene seas and accumulated outside the "Blue Earth" facies as well. Beyond the upper Eocene deposits, amber occurs in the terrigenous Oligocene and, in a redeposited state, in the Neogene to Quaternary sediments.

Abundant faunal remains are associated with the "Krant" facies, in which Noetling (1885) found 152 species of different organisms, from bryozoans to vertebrates. This fauna served as the basis for correlation of the Prussian Formation with the Lattorfian of northern Germany, where this subdivision is referred to the lower Oligocene. Abundant microphytoplankton of the *Hystrichosphaera* assemblage distinguished by Zatulka (1973) at the base of the Prussian Formation is rather similar to the late Eocene microflora of Ukraine. The bivalve assemblage with abundant *Cubitostrea plicata* Sol. that is typical of the Bartonian and Lattorfian (Kaplan *et al.*, 1977) was discovered in clayballs at the top of the "Blue Earth." On the basis of sponge spicules, the Prussian Formation is correlated with the upper Eocene of the middle Dnieper region, and dinoflagellate remains suggest its correspondence to the Priabonian Stage. The palynological assemblage from the "Blue Earth" and "Quicksand" is close in composition to those known from the lower part of the Kharkov Formation in Belarus, Obukhov Formation in the Dnieper region near Kiev, and Al'ma Horizon in the Crimea and Kuban' section of northern Caucasus (Grigyalis *et al.*, 1988; Didkovskii *et al.*, 1979). The remains of green algae, silicoflagellates, and diatoms were also encountered in sediments of the Prussian For-

Table 1. Distribution of foraminifers in the Prussian Formation and underlying clay bed of the Alk Formation

Beds of the Prussian Formation	Clay		"Wild Earth"		"Blue Earth"		"Quicksand"		"White Wall"
samples	1	2	3	4	5	6	7	8	9
Globigerina corpulenta Subb.	+		+		+	+			
Globigerina ex. gr. *danvillensis* Howe et Wall.	+			+	+	+			
Globigerina bulloides d'Orb.		+	+					+	
Globigerina praebulloides Blow	+		+	+					
Globigerina cf. *G. pachyderma* sin. (Ehr.)	+			+	+				
Globigerina eocaena (Gumbel)	+	+	+	+	+	+			
Globigerina sp. aff. *G. quinqueloba* Natland	+	+		+	+				+
Globigerina postcretacea Mjatl.	+			+	+				
Globigerina officinalis Subb.	+	+	+		+				+
Globigerina officinalis Subb. cf. *G. glutinata* (Egger)			+	+	+				
Turborotalia centralis (Cushm. et Bermudez)	+			+	+	+			
Globigerina turkmenica Khal	+	+							
Turborotalia cunialensis (Toum. et Bolli)					+				
Globigerina gortanii (Borsetti)			+						
Truncarotaloides cf. *rohri* Bronn. et Bermudez			+						
Globigerina sp.			+						
Spiroloculina sp.		+							

mation. According to the lithologic composition and fauna, similar to that of the stratotype, the Prussian Formation has counterparts in southwestern Lithuania, where they are represented by inequigranular glauconite–quartz sands barren of amber fragments and phosphorite concretions. Absence of these minerals most likely indicates that only upper units of the Prussian Formation, the "Quicksand" and "White Wall," accumulated here, because in earlier period this region represented an area of erosion adjacent to the Mazury–Belarusian uplift.

Foraminifers from sediments of the Prussian Formation of the Sambiiskii region correspond to the zones P15, P16, and P17 (to a part of the latter). Planktonic species *Globoquadrina corpulenta* (Subb.) found in concurrent deposits of Lithuania confirm this inference. The associated foraminifers are represented here by benthic forms (Grigyalis *et al.*, 1971). Benthic species are very characteristic of the upper Eocene sediments in Grodno and Gomel' areas of Belarus (Fursenko and Fursenko, 1961), and of the amber-bearing Potczyno Member in the Chlapowo region of northeastern Poland (Ordziwolska-Bienkowa, 1987). Southward, in the eastern lowland of Poland, the foraminifer assemblage from the upper Eocene marly sediments is more diverse and includes more planktonic forms (Pozaryska, 1977).

Our investigations in the Sambiiskii Peninsula showed that planktonic foraminifers prevail in the Prussian deposits from the Primorskii quarry of the Amber factory. Their assemblage includes *Globoquadrina corpulenta* (Subb.) encountered in the upper Eocene sediments of Lithuania, *Turborotalia centralis* (Cushman et Bermudez), *Truncorotaloides* cf. *rohri* Bronnimann et Bermudez, and *Globigerina praebulloides* Blow known from correlative deposits of eastern Poland. Three planktonic species of recent appearance were likely redeposited from younger sediments overlying the "Blue Earth" in the quarry. We did not find foraminifer species in common with those from the upper Eocene assemblage of Belarus. The majority of planktonic foraminifers represents small boreal forms, though the larger thermophilic taxa, such as *Turborotalia centralis* (Cushman et Bermudez), *Turborotalia* cf. *cunialensis* (Toumarckine et Bolli)?, *Truncorotaloides* cf. *rohri* Bronnimann et Bermudez, *Globigerina* cf. *gortanii* Borsetti, and *Globigerina* sp., were found as well. Planktonic foraminifers are most diverse in the upper part of the "Wild Earth" and in the lower part of the "Blue Earth" (Table 1). In general, the planktonic foraminiferal assemblage from the Sambiiskii Peninsula suggests that it inhabited a boreal sea with a relatively low salinity, which experienced from time to time the influence of warm and saline Tethyan waters. In the "Quicksand" and "White Wall" unit, foraminifers are very scarce that indicates the basin shoaling in corresponding periods.

In the Plyazhevyi quarry (see inset map to the figure) located about 3 km northward of the Primorskii quarry, the Prussian sediments are barren of foraminifers, suggesting most likely that water salinity was low-

Table 2. Correlation scheme of local stratigraphic subdivisions in Eocene–Oligocene amber-bearing sediments of Russia (Sambiiskii Peninsula), Poland, Lithuania, Belarus, and Ukraine

Era	System	Series	Stage	Planktonic foraminifer zones	Northeastern Poland		Sambiiskii Peninsula	Southwestern Lithuania	Northwestern Belarus	Northern Ukraine	
Cenozoic	Oligocene	lower	Stampian		Verkhnee Mosino Formation		Pal'vek Formation		Kharkov Group		Mezhgor'e Formation
					Shchepino Formation						
	Eocene	upper	Priabonian	P17	Nizhnee Mosino Formation	Chlapowo Subformation	Prussian Formation				Obukhov Formation
						Potczyno Subformation					
				P16	Pomeranian Formation	Upper Meroshino Subformation					
				P15							
		middle	Bartonian	P14		Lower Meroshino Subformation	Alk Formation		Kiev Formation		

ered near the Paleo-Neva mouth. Distinctions between foraminiferal assemblages from different areas of the Late Eocene strait are connected with variations in facies composition of deposits: predominance of benthic species is indicative of shallower sedimentation settings, and planktonic forms are typical of deeper ones. Predominance of planktonic forms in the assemblage from the Sambiiskii Peninsula suggests that the basin was about 200 m deep.

DISCUSSION AND CONCLUSION

Sediments considered above, those of the amber-bearing Prussian Formation included, accumulated under unstable tectonic conditions, when the young fold systems of Middle Europe developed between the East and West European platforms. Environments of sedimentation were determined here by the Danish-Poland aulacogen that completed its development above the large Teisseira–Tornquist lineament at that time (Khain, 1977; *Geologicheskie i bioticheskie sobytiya...*, 1996).

Paleogeographic environments during the amber accumulation were under control of the marine strait that connected the tropical Tethys with the North Atlantic and Boreal oceans, representing the seaway for water exchange (Kharin, 1997). This is evident from coexistence of thermophilic and cold-resistant planktonic foraminifers in sediments of the Prussian Formation. The amber-bearing Prussian deposits accumulated owing to transportation of terrigenous material and resin from northern regions (central and northern Baltic Sea areas and Scandinavia), where marshy landscapes were occupied by coniferous "amber" forests. Strong monsoon winds damaged trees, which oozed resin that was transported by a large river, Paleo-Neva, to the lagoon system in the northwestern Kaliningrad region, where primary amber deposits accumulated. Subsequent erosion of the latter and redeposition of amber led to the formation of secondary, placer deposits confined to different sedimentary facies.

The taxonomic similarity of fauna and flora from amber-bearing sediments of Belarus, Ukraine, Poland, and northern Germany and in those of the Sambiiskii Peninsula, which usually bear glauconite, and phosphorites, should indicate their concurrent accumulation. However, there is no agreement on the age of amber-bearing deposits. In Denmark, the amber-bearing Mo Formation is referred to the Middle Eocene (Katinas, 1987). Researchers from St. Petersburg (Kaplan *et al.*, 1977) estimated that "Blue Earth" of the Sambiiskii Peninsula accumulated about 37.7 Ma ago, i.e., in the late Eocene. The same age is inferred for the "Wild Earth" and "Blue Earth" beds based on investigation of their spores, pollen, and phytoplankton (Kosmowska-Ceranowich *et al.*, 1997). On the other hand, the K–Ar age determined by Ritzkowsky (1997) for the "Blue Earth" is 44.1 Ma that corresponds to the middle Eocene. According to nannoplankton, the amber-bearing Potczyno Member near Chlapowo is referred to the

zone NP21 corresponding to the upper Eocene–lower Oligocene boundary, whereas the foraminiferal fauna implies the uppermost Eocene level of this unit (Ordziwolska-Bienkowa, 1987; Kosmowska-Ceranowich and Muller, 1985). In Ukraine, the amber-bearing Mezhgor'e Formation is defined to be of the early Oligocene age (*Stratigraficheskaya skhema...*, 1987), while in Belarus the amber-bearing sediments are attributed to the Kharkov Series of the upper Eocene–lower Oligocene (Basharkevich *et al.*, 1983). These disagreements are likely to be objective, understandable from the standpoint of facies law by Golovkinskii and Walter that controlled the facies migration in unstable tectonic and paleogeographic settings in the sedimentary basins associated with the aforementioned marine strait and Danish–Poland aulacogen. The issue remains still open, requiring a further clarification. The absolute age determinations for glauconite from amber-bearing deposits are also problematic, because this mineral may be of different origin, either redeposited from Upper Cretaceous, Paleocene, and Eocene sediments, or syngenetic, formed owing to transformation of feldspar and volcanic ash. The ash could be transported to the amber accumulation site from volcanoes of the Skagerrak Strait and northern Germany (*Atlas paleogeograficheskikh kart...*, 1993; Jacque and Thouvenin, 1975).

Findings of amber and phosphorite in Paleogene sediments of Belarus, Ukraine, Poland, Denmark, Germany, Lithuania, and Russia can be used in addition to analysis of fauna and flora, in correlation of local stratigraphic units (Table 2). Investigations of this kind are of practical importance for the geologic mapping and prospecting of amber and phosphorite deposits in the Paleogene sediments.

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The Tarkhanian–Karaganian Stage in Development of the Euxinian–Caspian Basin (Eastern Paratethys). Part 2[1]

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Abstract—Evolution and confinement of biota to concrete sea zones, and paleogeography of the Euxinian–Caspian basin (the Eastern Paratethys since the initial Oligocene) at the beginning of the middle Miocene (Chokrakian) are considered. It is shown that this was a time of most intense structural reorganizations in the basin. The changes were related to the growing differentiation of the sea floor: to shoaling and partial exsiccation of an axial part and to subsidence of shelf depressions. Surrounding land areas, which grew in size at the same time, experienced orogenic movement in the south as a consequence of collision between the Anatolian–Iranian and Afro-Arabian continental blocks. Relationships of the initial mid-Miocene time between the Euxinian–Caspian, Central Iranian, and East Mediterranean basins are discussed.

Key words: Eastern Paratethys, Chokrakian, Karaganian, paleogeography, tectono-sedimentary cycle, biota evolution, transgression, regression.

In the Chokrakian time, floor of the basin experienced an intense differentiation, and this event is correlative with the Styrian orogeny identified by G. Stille (Shcherba, 1993; see table and corresponding figures in Part 1). In parallel with subsidence of deep shelf zones, undersea terraces turned out to be relatively uplifted, partially above the sea level. As a consequence of synfolding uplifts (cordilleras) in the axial part, the basin narrowed. The event activated accumulation of sediments in the axial and shelf depressions, causing their shoaling and intensifying the avalanche sedimentation, land sliding and slope erosion (Figs. 2 and 3). The concurrent downwarping of regional troughs along the northern and eastern margins of the basin led, in contrast, to the basin widening.

The early and late Chokrakian stages in the basin evolution are easily distinguishable. After cessation of connections with the Western Paratethys (Goncharova, 1989; Goncharova and Il'ina, 1997; Goncharova and Shcherba, 1997), the early Chokrakian basin was left connected, via the Middle Araks strait (Fig. 2, *18*), with the Central Iranian basin that was already separated from the Indo-Pacific part of the Tethys (Adams *et al.*, 1983; Gansser, 1986; Steininger *et al.*, 1985) and advanced to the north. It is likely that this and Southern Van (between the Van and Urmia lakes) straits supplied the Chokrakian basin with East Mediterranean and Indo-Pacific waters. The Southern Van strait connected the Central Iranian basin with the northern part of the Mesopotamian basin that was open to Indian Ocean in the southeast [basal mid-Miocene carbonate sediments with abundant marine fauna in the Diyarbakir depression of southeastern Anatolia and northeastern Syria (Čtyroky *et al.*, 1975; Ponikarov *et al.*, 1969; Krasheninnikov *et al.*, 1999)]. Northwestward of the Arabian plate, the Mesopotamian basin was also connected with the East Mediterranean sea (Ponikarov *et al.*, 1969) via the Aleppo–Urfa seaway that consisted of shallow distributaries in the west. The northern seaway arm, sediments of which bear abundant terrigenous admixture (Urfa–Adjaman–Gaziantep zone, Steininger *et al.*, 1985) was presumably open to the Adana bay that accumulated Langhian carbonate deposits (Steininger *et al.*, 1985, Görür, 1994; Williams *et al.*, 1995; Yetis *et al.*, 1995). Marine deposits of the basal middle Miocene are also known in the Antioch and Latakia depressions (Erünal-Erentöz, 1950; Ponikarov *et al.*, 1969; Krasheninnikov, 1971; Krasheninnikov *et al.*, 1999), which probably mark the former positioning of southerly seaway arms.

Influence of the Middle Araks strait that connected the Early Chokrakian Basin with adjacent seas is evident from the following observations: diversity of Chokrakian bivalves and gastropods increases from the northwest to southeast; migrant genera and species that are absent from the Tarkhanian and Mediterranean basins are very abundant in the Transcaspian region (Goncharova, 1989; Studencka *et al.*, 1998; Il'ina, 1993, 1995, 2000); assemblages of planktonic foramin-

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ifers from Azerbaijan are of peculiar composition (Mamedova, 1974). Deposits that accumulated in the southeastern part of the Middle Araks strait (northwestern Iran, Dash-e-Moghan area in the Araks River lower courses) yield abundant Chokrakian fauna. This fauna includes Mediterranean bivalve species *Chamelea fasciculata* (Reuss) and *Divalinga ornata* (Agassiz), which inhabited the South Anatolian and, probably, Central Iranian and Mesopotamian basins, but were untypical of the Chokrakian sea (Gillet, 1967). We should also note the peculiar composition of Chokrakian mollusks from the Dzhebrail bay (northwestern part of the Middle Araks strait), which include Mediterranean species *Mactra basteroti, Lunatia helicina*, and five *Turritella* forms (Alizade *et al.*, 1980), all untypical of the Chokrakian molluscan fauna proper.

The increased area of the early Chokrakian shallow-water settings favored flourishing of benthic mollusks, bryozoans, foraminifers, and ostracodes. Bivalves (70 species, 31 of which are of the Badenian and/or Mediterranean origin) were mostly inherited from their previous basin. Migrants represented less than one third of the population whose endemism increased by 10–11% so that some new species were characteristic of the early Chokrakian only (Goncharova, 1989). Bryozoans (35 species of Cyclostomida and Cheilostomida orders, the latter dominant) were abnormally diverse and abundant (Veis, 1988). Miliolids prevailed among benthic foraminifers, though polymorphinids and nonionids were also rather abundant (Bogdanovich, 1965). Ostracodes of increased diversity included representatives of euryhaline and polyhaline genera (*Neogenovaya sistema*, 1986b). Since the time of N.I. Andrusov, the diversity and abundance of the early Chokrakian fauna attracted attention of researchers to these peculiar fossils.

Most diverse communities populated shallow-water settings, which were not deeper than few tens of meters, well aerated, and rich in nutrients. These settings hosted individual bioherms and their chains, which rose above the bottom. The oval biohermal bodies (up to 2 m across) are composed of bryozoan colonies (dominant genera *Membranipora* and *Schizoporella*, Veis, 1988), red algae (Belokrys, 1986), and calcareous worm tubes. Proportions of frame-building taxa vary along some chains of bioherms, but bryozoan and algae are usually dominants. Mollusks concentrated in peripheral zones of bioherms are often covered with dense calcareous crust that perfectly preserves their shells and is easily removable. Genera *Pteria* and *Chama* are dominant among bivalves (about 25 species in total). Representatives of genera *Parvicardium, Anadara, Venerupis, Irus, Gregariella*, and *Mytilaster* are also abundant and occur in somewhat different proportions in various areas (e.g., in Crimea, Northern Caucasus, and Georgia) of the Chokrakian basin (Goncharova, 1989). Pectinids (genus *Aequipecten*) and representatives of genera *Mytilus, Isognomon, Ervilia, Corbula*, and *Hiatella* are not as abundant in bioherms as the aforementioned taxa. Species of *Nuculana* and *Modiolus* genera are rare. Species *Cerithium catleyae* Baily, *Bittium (B.) digitatum* (Zhizh.), *B. (?) agibelicum* (Zhizh.) and *Potamides (Pirenella) cicur* (Zhizh.) are most abundant among gastropods (field observations of 1973 by Bagdasaryan, determinations by Il'ina, 1993). Their species *Nassarius (Hinia) restitutianus* (Font.), *Gibbula (Colliculus) kertschensis* (Usp.), and *G. (C.) nefas* Kol. are less frequently occurring. Representatives of Balanidae, Polychaeta (*Serpula* and *Spirorbis (Dexioospira) fundatus* Belokrys) are known overgrowing coquina and bioherms of the Kerchenskii Peninsula and coquina in the Ustyurt. In their biota composition, the Chokrakian bioherms seem to be close to facies of biohermal limestone distinguished by Studencka (1999) in the Badenian horizon of Rhodophyceae Limestone in the Carpathian foredeep of Poland (Western Paratethys). Fauna from Poland is however more diverse than that from the Chokrakian bioherms.

Most abundant among bivalves, which populated coquina-sandy grounds of the upper subtidal zone, were mitilids and arcids, whereas representatives of *Gibbula* and *Cerithium* genera were most frequent among gastropods (Bagdasaryan, 1965; Goncharova, 1989). Bryozoans and Balanidae were also abundant here. Benthic foraminifers were mostly represented by rotaliid species *Streblus beccarii* (L.) (Bogdanovich, 1965).

Community of sandy grounds widespread in shoal zones was dominated by bivalves of the Arcidae, Pectinidae, Donacidae, Veneridae, Mactridae, and Mesodesmatidae families (Goncharova, 1989). *Gibbula* and *Cerithium* forms, in association with bryozoans and *Spirorbis* species, were also widespread in this ground facies. In the upper and intermediate subtidal zones, grounds composed of fine-grained sand were populated predominantly by polymorphinids and miliolids (genus *Tschokrakella*), which thick-walled shells with coarse costae and often coexisted with *Elphidium rugosum atschiensis* Bogd. (Bogdanovich, 1947, 1965). Ostracodes were extremely diverse in these settings. Representatives of Polychaeta (genus *Ditrupa*) were abundant at the end of the early Chokrakian in some desalinated areas with fine-grained sandy grounds. Lenticular accumulations of their tubes occur in western Georgia (Dzhgali district, Bagdasaryan, 1965) and Kerchenskii Peninsula (western area of the Malyi Kamyshlak Ravine). Shoal areas floored by muddy sand were occupied by the *Ostrea digitalina* (Eichwald)–*Aequipecten varnensis* (Toula) community that included echinoids of the genus *Schizaster* (Georgia and Transcaspian region).

In the lower and intermediate subtidal zones, polymorphinids forced out miliolids from fine-grained sandy grounds to become dominating there (Bogdanovich, 1965). Bottom areas of the lower subtidal zone, which were covered with silty mud, hosted *Nuculana (Saccella) fragilis, Corbula (Varicorbula) gibba,*

Macoma sokolovi, Cultellus paryraceus, Nuculana (Lembulus) emarginata, and *Lutetia (Davidaschvilia) intermedia*. Miliolids and polimorphinids were widespread in muddy areas of the same zone, but nonionids, in particular, *Florilus boueanus* (Orb.), were less abundant here.

The miliolid assemblage with *Discorbis tschokrakensis* Bogd. (Discorbiidae) was widespread above muddy grounds of the lower subtidal zone (Bogdanovich, 1947, 1965). Pure mud at the pseudoabyssal depth represented habitat of miliolids, the *Tschokrakella* species with fragile shells included, which associated with abundant nonionid species *Florilus boueanus* (Orb.), ophiurids, and thin-walled bivalve species *Nuculana subfragilis, Abra parabilis*, and *Thyasira flexuosa*.

Plankton was dominated by *Spiratella* forms, though they were less abundant than before. Pelagic zone of the Early Chokrakian basin was predominantly inhabited by Gadidae and Clupeidae, one third of which or more were inherited from fish community of the Tarkhanian time, and by newly appeared Scombridae. Desalination and shoaling of the basin caused disappearance of deep-water *Vinciqueria merklini* Dan. (*Neogenovaya sistema*, 1986b).

Owing to connections with southern seas, salinity of the Early Chokrakian basin somewhat deviated from the normal one, but did not decrease below 28‰ (Bagdasaryan, 1959, 1965) except for local water masses, which were under influence of river runoff. This is evident from occurrence of echinoids, spiratells, and hexacorals (single find, determination by M.R. Gekker), and also from taxonomic composition of bivalves (Goncharova, 1989) and bryozoans (Veis, 1988). Thus, salinity of the Early Chokrakian basin was higher than that suggested by Zhizhchenko (1940) and Manolov (1960).

In the late Chokrakian time, after isolation in the southeast, the basin considerably grew in size, mostly in the European part of its shelf zone, owing to the intense fresh-water influx and shoaling of depressions filled with sediments. Isolation from the global ocean resulted from partial or complete closure of the Middle Araks strait that periodically connected the Eastern Paratethys with Southern Peritethys during the Neogene time. Hydrological regime of the basin turned into semimarine (Goncharova, 1989), and the water salinity approached that of the subsequent Karaganian basin (Bagdasaryan, 1959).[2] This event impoverished fauna of the basin, and some taxa became extinct, e.g., 6 bivalve species (Goncharova, 1989) and 7 species of benthic foraminifers,[3] which are dwarfed, showing abnormal evolutionary trends (Bogdanovich, 1950; *Neogenovaya sistema*, 1986b). Dwarfing was also typical of rather abundant ostracodes (Bogdanovich, 1950). The fauna was lacking planktonic foraminifers and spiratells. Only Clupeidae survived among fishes. Presence of gastropods is indicated by rarely occurring embryonic shells of Prosobranchia (reference to Il'ina in Goncharova, 1989; Il'ina, 1993), which were mistakenly identified as *Spiratella* forms in many cases.[4] Not more than one to three bivalve species were able to transit in the subsequent Karaganian basin (Goncharova, 1989) along with single representatives of Diastomidae and Nassariidae (Il'ina, 1993). Many assemblages of late Chokrakian fossils include abundant specimens of Polychaeta, for instance, "*Spirorbis (Dexioospira)*" *obliquinulatus* Belokrys known from the Kerchenskii Peninsula and some other Black Sea regions. In the late Chokrakian time, these organisms dwelt over the blue-green algae and together with them participated in construction of stromatolites (Belokrys, 1984). Genus *Uruchia* that became widespread in the Karaganian basin was most important among stromatolitic algae, which populated steep undersea slopes (Kholodov and Rodionova, 1979). *Pectinariopsis* tubes (Polychaeta) are widespread in the central Cis-Caucasian region. They are built of small *Lutetia* and *Ditrupa* shells in the upper Chokrakian deposits, and of young *Spaniodontella* shells in the Karaganian beds.

In the early Chokrakian time (Fig. 2), the axial part of the basin became almost twice narrower as a result of contraction of the Lazarevskoe–Kobystan and Adzharia–Trialeti troughs, the deep-water structures of the Paleogene marginal sea of the Greater Caucasus (Fig. 2: 3). A relic of the continental slope break that is preserved in the Tengiz–Beshbarmak anticlinorium of southeastern Caucasus represented a shoal area advanced to the south. This area periodically rose above sea level, and triggered the land sliding that resulted in deposition of dolomite breccias with plant remains (the petrified *Dsuduba* trunk, Baikovskaya, 1950) buried in clay near the slope base. In the northwestern Caucasus, the entire area of continental slope, except, probably, for its basal part, was risen up to the shallow shelf level, as it is evident in the Sochi district, where clayey sediments of a deep shelf zone are exposed. The shelf was rimmed by reefal buildups, which are traceable in the present-day relief of the Anapa suburbs and within an area extending from the Benokovo (Laba River) and Abadzekhskaya (Belaya River) westward to the Khadyzhensk and Kaluzhskaya village (Shcherik, 1973; Dmitrieva, 1953). The middle part of this slope was likely transformed into the Goitkh synfolding cordillera that periodically rose above the sea level.

In the central Caucasian transect, the Lazarevskoe–Kobystan trough became closed completely. An insta-

[2] Analyzing composition of foraminiferal assemblages from the West Kuban' depression, Pinchuk in Bigun *et al.* (1998) concluded that salinity of Late Chokrakian basin was about 20–18‰.

[3] According to data of T.N. Pinchuk (personal communication), the late Chokrakian benthic foraminifers are represented by 18 species, mostly by miliolids.

[4] Presence of spiratells in upper Chokrakian deposits should be verified, because representatives of the genus *Limacina* (taxon of present-day pteropods identical to genus *Spiratella* in paleontological taxonomy, Janssen, 1989) can survive the water salinity not lower that 28‰ (Ivanova, 1983).

ble land area from the south that appeared here served as a bridge for African and Eurasian mammals, which migrated inside the basin area (Fig. 3). It was a time, when aardvarks (genus *Orycteropus*) and Suiformes (genus *Kubanochoerus*), the representatives of Belomechetka fauna from the Stavropol' uplift, appeared in the Caucasus together with Eurasian bears and deer. At the same time, the endemic Caucasian genus *Paranchiteria* migrated to the Balkan Peninsula (Goncharova and Shcherba, 1997; Popov *et al.*, 1998, with reference to V.I. Zhegallo). The land appearance instead of the former deep-water basin is also evidenced by occurrence of leaf flora in the northern Kartli area (Akhmetiev, 1995). Vegetation migrated there most likely from the north.

Deep-water settings were preserved only in the east of the Lazarevskoe–Kobystan trough and in the Kobystan–South Caspian depression (Figs. 2 and 3: 1). The northern half of the latter accumulated either the condensed carbonate-free barren deposits, or pseudoabyssal mud sequence (150–200 m thick) that yields *Spiratella* forms and impoverished assemblage of benthic mollusks (*Neogenovaya sistema*, 1986b). An extreme scarcity of benthic foraminifers here reflects most likely a stagnant state of bottom and near-bottom water layers contaminated by hydrogen sulphide, as it is evident from abundance of synsedimentary pyrite crystals (Mamedova, 1974). Closer to the Alyat Ridge, mud was calcareous, mixed with sandy fluxoturbidites, which increase threefold the total thickness of Chokrakian deposits at the southern slope of the trough (Kopp and Shcherba, 1985). Despite the presence of *Spiratella* forms, break of this slope was located not deeper than the lower subtidal zone, as it is indicated by occurrence of the following faunal groups: benthic mollusks of genera *Anadara, Ervilia*, and *Cerithium*; benthic foraminifers of genera *Articulina, Florilus*, and *Quinqueloculina*; ostracodes of genera *Loxoconcha, Leptocythere*, and *Cytherura*; and otoliths representing bullheads of the genus *Gobius* and Percoidea genus *Percidarum* (*Neogenovaya sistema*, 1986b; Mamedova, 1974). Abundance of this relatively shallow-water fauna of the early Chokrakian time points to the normal aeration, only slightly changed salinity, and high temperature of seawater (Mamedova, 1974). In response to the late Chokrakian cessation of connection with open sea, this part of the trough turned first into a lagoon that was most deep in the Tarkhanian, and then into a land area, where continental sediments with conglomerate interlayers were deposited. Only at the time of the Karaganian transgression, this area was flooded by seawater and turned into an upper subtidal setting, where spaniodontells were dominant habitants (Alizade *et al.*, 1980).

Undivided deep-water deposits of the middle Miocene, which accumulated in the East Black Sea and Adzharia–Trialeti depressions that decreased in size (its part near Tbilisi turned into the terraced shelf, Figs. 2: 3), occur northward of the Natanebi thrust fault, near which the monotonous calcareous clay of this age was penetrated by boreholes (Tugolesov *et al.*, 1985). The analogous clay sequence that is 80 to 150 m thick and yields *Nuculana subfragilis, Thyasira flexuosa*, and *Spiratella* specimens from its lower part and *Lutetia intermedia* from the upper one (Bagdasaryan, 1965) was also discovered in the southern Kolkhida lowland (between Khobi and Chanis-Tskhali rivers, and near the town of Tskhakaya at the southern slope of the Tsaish Ridge).

The southern part of the Adzharia–Trialeti depression was intensely filled with terrigenous material derived from the Lesser Caucasus land during the Chokrakian time (Fig. 3, profile I–I). Sediments, which accumulated during the middle Miocene in the Guriya depression (Fig. 2: 2), are twice thicker than concurrent sediments in other areas of the East Black Sea depression (more than 3 versus 1.5 km, Tugolesov *et al.*, 1985). In the northern part of the Guriya trough, outside the present-day contours of the Black Sea, the Tarkhanian–Chokrakian interval corresponds to the sequence of *Spirialis* Clay that is up to 500 m thick and encloses subordinate interlayers of bituminous clay, marl, and clayey sandstone, which bear an impoverished fauna of the pseudoabyssal type (Maslov, 1937; Zhizhchenko, 1940). Closer to the southern wall of the trough, where clay beds leaned against a steep flexure inside the Paleocene–Eocene deposits, the deep-water sediments enclose sand beds with leaf flora (Palibin, 1936) and local accumulations of diverse early Chokrakian fauna. By the end of the Chokrakian, sedimentation developed here in supralittoral environments (Zhizhchenko, 1940) and resulted in deposition of coarse sand and conglomerate beds (130–250 m, Bagdasaryan, 1965). The greater part of Adzharia–Trialeti shelf was transformed into land periodically flooded by sea. The eastern deep part of Adzharia–Trialeti trough apparently was closed in response to the tectonic contraction, but this does not exclude completely the influence of sedimentation factor.

In general, the middle Miocene evolutionary stage of East Black Sea depression and Guriya depression was characterized by their gradual sedimentary compensation that transformed the former abyssal plain into pseudoabyssal. In the East Black Sea depression, northern slope was less steep than during the Paleogene time, but the depth gradient here was still considerable. The West and East Black Sea depressions joined each other smoothly, but sedimentation in the former that remained deep till the end of the Chokrakian was 1.5–2 times more intense than in the latter (Tugolesov *et al.*, 1985).

The Fore-Talysh (Astrakhanbazar) depression (Figs. 2 and 3: 4) that was about 200 m deep, representing a pseudoabyssal zone during the early–middle Tarkhanian, was filled during the late Tarkhanian–Chokrakian time with sandy–clayey sediments, which are 500–600 m thick in its northeastern part. The north-

ern part of the trough (not deeper than 200 m) hosted an assemblage of microfauna that impoverished with time. Plant remains and volcanic ash trapped here were derived from the volcanic plateau of the Lesser Caucasus. By the Karaganian, this part of the trough shoaled, and calcareous ooze became a noticeable component in its filling-in (Alizade *et al.*, 1980).

Thus, a considerable part of axial zone of the Greater Caucasus basin represented in the Chokrakian time a shelf sea coalesced with northern and southern shelf areas of the Eastern Paratethys. In its depth range, the sea corresponded to lower and upper subtidal zones. Shoaling of undersea swells, which separated deep-water troughs, was most noticeable. In response to expansion of the Caucasian land, the shallowest area appeared in the east of the Schatsky swell (Fig. 2: 9), above the Dzirul massif (basement of the Rioni and Kartli intermontane depressions). Sandy deposits of the upper subtidal zone, which accumulated along the periphery of the massif, yield the diverse early Chokrakian fauna of fossil mollusks, benthic foraminifers, and echinoid species of the genus *Schizaster* (Bagdasaryan, 1965).

During the Chokrakian, the Tuapse and Shemakha troughs (Fig. 2: 6 and 7) trapped terrigenous material as effectively as in the Oligocene that is clearly visible in seismic profiles (Tugolesov *et al.*, 1985). Subsidence of the Tuapse depression that accumulated mud with poor monotonous fauna of *Nuculana fragilis* and *Abra parabilis* (Bagdasaryan, 1965) resumed in the initial Chokrakian time, when its southern slope experienced erosion and concurrent land-sliding events (Borukaev *et al.*, 1981). Under conditions of suddenly increased sedimentation rate, both troughs turned out to be almost filled in, by the end of the Karaganian, with the 100-m-thick sequence of clay and sand with lignite lentils.

During the Chokrakian, the Iori–Kura depression in the southern shelf (Fig. 2: 5) was a principal site of terrigenous material discharge. This deep (lower subtidal–(?) pseudoabyssal zone up to 150–200 m deep) sedimentary trap accumulated mud (sequence up to 900 m thick) until the mid-Sarmatian time (Shcherba, 1993). Sandy material with pebbly admixture and abundant plant remains (Akhmetiev, 1995) periodically accumulated in the western slope of the trough (Norio–Martkobi area). This material was derived from the uplift of the Lesser Caucasus (Grossgeim and Gmid, 1951), along which a belt of supratidal deposits (50–250 m) with diverse benthic fauna (Zhizhchenko, 1940; Bagdasaryan, 1965) marks the former position of shoreline that retreated southward during the late Chokrakian–Karaganian transgression. Outliers of Karaganian coastal-marine coquina, sand, and clay are preserved in the Lesser Caucasus in areas with altitude of 700–1100 m (Milanovskii and Khain, 1963).

Conglomerates of the near-shore zone also surround the northwestern Talysh (Fig. 3, Alizade *et al.*, 1980). In the Araks River middle courses, the shoreline deviated however to the south and delineated a narrow strait that separated during the Tarkhanian–early Chokrakian period the Lesser Caucasus and Elburs–Kopetdag rises and connected the southern part of the Chokrakian basin with the Central Iranian (Khain, 1984, with reference to Berberyan and King) and East Mediterranean basins (Goncharova and Shcherba, 1997; Studencka *et al.*, 1998). Near-shore carbonate sediments that accumulated in the southeast of the strait (Dash-e-Moghan site in northwestern Iran) contain the Chokrakian assemblage of diverse benthic fossils (Gillet, 1967), many of which are of Mediterranean origin. The Dzhebrail trough in the northwest of the Middle Araks strait represented a sedimentary trap in the southeastern slope of the Lesser Caucasus, where the entire Miocene sequence is over 1200 m thick (Alizade *et al.*, 1980). In the Tarkhanian, this trough accumulated carbonate mud of the relatively deep-water type (deep-shelf sediments). These sediments grade upward into the topmost Tarkhanian(?)–lower Chokrakian rudaceous deposits with fragments of shallow-water *Ostrea* Limestone that bears the diverse Chokrakian fauna, Mediterranean migrants included, which were redeposited by debris flows. The avalanche sedimentation quickly filled in the trough, and shallow-water sands with fucoids and exsiccation cracks accumulated here already by the end of the Chokrakian time. During the Karaganian, the Dzhebrail trough area rose above the sea level (Alizade *et al.*, 1980). In general, it is very likely that area under consideration was isolated from the Central Iranian and East Mediterranean basin before the beginning of the late Chokrakian time (Goncharova and Shcherba, 1997), as it is evident from the mass extinction of marine fauna.

Geological transformations triggered by the Styrian orogeny were even more distinct in the **Iran–Lesser Caucasus** structural zone that framed the Euxinian–Caspian basin in the south. Orogenic movements extended area of the Lesser Caucasus (Somkhet–Agdam) lowland (Volcanic plateau of the Kura River head) that was bounded on the north by the aforementioned belt of near-shore marine deposits. At the same time, the Central Iranian sea basin that lost connections with the Indo-Pacific ocean advanced northward, into the eastern Anatolia, as the South-Armenia gulf (Fig. 2: 16). This gulf reached the northeastern coast of the present-day Lake Van (Gelati, 1975; Bossiouni, 1979; Steininger *et al.*, 1985; Sengör *et al.*, 1985) and extended to 40.5°N. Its northeastern ending corresponded to the Sevan–Aragats lagoon (Fig. 2: 17). According to Arzumanyan (1980), the northwestern part of the latter (junction site of the Akhuryana and Araks rivers) experienced periodical flooding, whereas central and eastern parts (Erevan and Shirak depressions, and Sevan graben-syncline) represented the accumulation sites of halite and gypsum (lagoonal deposits from 40 to 500 m thick). The western part of the lagoon (westward of Erevan) corresponded to a zone of intense volcanic activity (liparite ash and tuffs)

characteristic of the Central Anatolian region (*Neogenovaya sistema*, 1986a). The eastern part of the South-Armenia gulf (Nakhichevan' area) with sandy to small-pebbly grounds was populated predominantly by benthic gastropods of the genus *Sandbergeria* (Alizade *et al.*, 1980, though Il'ina identifies it as the Indo-Pacific genus *Obtortio*), which associated with scaphopods of the genus *Dentalium*. Deeper basin areas on the west (42–100 m deep according to Alizade *et al.*, 1980) were floored by muddy sand and populated by planktonic and benthic foraminifers, and by pteropods of the genus *Spiratella*. The northwestern part of the South-Armenia gulf (area extending from the Lake Van to 40.5° N and to Erzerum in the west) periodically experienced exsiccation (Lüttig and Steffens, 1975). Gypsum-bearing molassic deposits between Erzerum and Sivas do not belong to the middle Miocene (Poisson *et al.*, 1997).

Area between Latakia and Adana, which was occupied during the middle Miocene by narrow arms and islands of the Aleppo–Urfa strait, was likely most favorable for migration of mammals. This very site was apparently used by African mammals who migrated northward and by Eurasian ones who advanced in opposite direction, because dwellers of both continents are present in the Belomechetka burial of the terminal early Chokrakian time (Popov *et al.*, 1998). In addition, *Gomphoterium* ex gr. *angustidens* (Cuv.), similar to its counterpart from this burial and widespread in lower–middle Miocene deposits of the Old World, is also known far away in the south, in the middle Miocene lacustrine deposits of Palmerides (Devyatkin *et al.*, 2000). The same deposits yield ostracode species *Cyprideis gigantea* Vorosh. (Ponikarov *et al.*, 1969; Devyatkin *et al.*, 2000) known from Karaganian deposits of the Nakhichevan' area, although representatives of the genus *Cyprideis* are unknown in pre-Karaganian strata in the Eastern Paratethys (*Neogenovaya sistema*, 1986b).

The northern shelf of the Euxinian–Caspian basin that lost its connection with the Western Paratethys experienced the most intense differentiation during the Chokrakian time. On the one hand, the West Kuban' (Fig. 2: 11) and, particularly, Terek-Caspian (Fig. 2: 12) depressions subsided again to the sediment compensation level (Shcherba, 1993). On the other hand, a considerable part of the northern shelf shoaled and was partially transformed into the land areas. These were the lowland of the Crimea Peninsula bounded in the southwest by shallow-water oyster banks, the Stavropol' rise and Laba–Malka zone of the northern Caucasus (Fig. 2: 14a), which joined together to form the Greater Caucasus island (Fig. 2: 14) that showed itself for the first time as an important provenance of terrigenous material (Kholodov and Nedumov, 1996), and the shallow East Kuban' terrace. The existence of a large island in the northern slope of the Greater Caucasus during the terminal early–late Chokrakian time is evident from the following observations: terrestrial mammals are components of the Belomechetka (Gabuniya and Benukidze, 1984) and Labinsk (V.I. Zhegallo, personal communication) burials; occurrences of terrestrial (chelicids) and fresh-water (unionids) mollusks are known at the Belaya River (Abadzekhskaya site); fossil insects are discovered near Stavropol' (Bekker-Migdisova, 1964; Dlusskii, 1981); the corresponding character of water currents is established in the southern Stavropol' district (Grossgeim, 1965); sediment transport into the West Kuban' depression was from the south (Dmitrieva, 1953) and southeast (Bigun *et al.*, 1998). Alluvial deposits to the south of Labinsk, which yield bone remains, apparently accumulated in delta of a river that drained the western side of that island. The river likely followed an older drainage system, because sandy debris cones accumulated in this area since the late Oligocene–initial Miocene time, and sandy material was derived from the incipient rise (Grossgeim, 1965; Shcherba, 1993). Development of a river system and remains of the early Chokrakian conifers from the western Cis-Caucasus (Burova, 1999; the middle Chokrakian in her stratigraphic scheme corresponds to the early Chokrakian in our interpretation) point to the mountain relief of the Greater Caucasus island that was earlier suggested by Dmitrieva (1953). In the late Chokrakian, near-shore settings on the northeastern side of the island were very diverse and left behind biocoenoses of lagoons, supratidal zones, fresh-water limans, reed marshes, and coastal shrubs (Bekker-Migdisova, 1964). This diversity and desalinating influence of river runoff was perceptible since the terminal early Chokrakian time, as it is evident from occurrence of *?Parvicardium brykense* Gontsh. et Zhgneti. This descendant of *Parvicardium kubanicum* (Zhizh.) that is similar in many features to the Sarmatian genus *Obsoletiformes* (Goncharova, 1989) illustrates an overdone evolutionary transformation of bivalves in the Chokrakian basin. The Greater Caucasus island, Dzirul land, and Lesser Caucasus could have periodical connections since the end of the early Chokrakian, which were favorable for migration of mammals from the south into the central Cis-Caucasian region, as it was originally suggested by Zhizhchenko (1940).

An archipelago of islands (shoals), which appeared in the eastern Caucasus, extended from the Greater Caucasus island as far to the east as the Apsheron Peninsula (Fig. 2: 15), i.e., for a distance of about 600 km. The archipelago is reconstructed on the basis of sediment mineralogy (Grossgeim and Gmid, 1951; Kholodov and Nedumov, 1981) and paleobotanical investigations, which discovered the leaf flora burials in the Terskii Ridge between Groznyi and Chir-Yurt (Rozanov, 1925), and in localities at the Sulak, Argun (Palibin, 1936; Akhmetiev, 1995), and Fortanga rivers (remains of mangrove vegetation in the last case, Kholodov *et al.*, 1980; Akhmetiev, 1995). The additional evidence is presence of the aforementioned large wood remains in the deep-water deposits of the Kobystan trough adjacent to the Tengizbeshbarmak

Island. In the southern part of the Terek–Caspian depression at the Derbent latitude, there was a minor undersea swell, the sandy to silty deposits of which yield mollusks of the upper subtidal zone (Grigorovich-Berezovskii, 1925).

The southern part of the Terek–Caspian depression represented deepest zone of the North Caucasian shelf. It was at the pseudoabyssal depth in the Tarkhanian time (Zhizhchenko, 1940) and turned into a deep shelf area in the Chokrakian. Small antithetic rises along its southern margin, i.e., islands of the aforementioned archipelago of the eastern Caucasus, were similar to those originated in the Oligocene time along the southern wall of the West Kuban' depression (Shcherba, 1993; Kopp and Shcherba, 1998) that corresponded to a deep shelf zone. The depressions were quickly filling in mainly by terrigenous material derived predominantly from a vast East European plain. In particular, the Kuma depression (Fig. 2: 13), except for its western part, was wholly compensated by sediments.

Shelf deposits, which accumulated in the northern marginal zone of the Chokrakian basin (Cis-Caucasian region), are characterized by an irregular distribution of coarse sandy deposits that have been studied by many researchers (Golynets, 1932; Zhizhchenko, 1940; Grossgeim and Gmid, 1951; Kosarev *et al.*, 1965; Buryak *et al.*, 1968; Kosarev, 1971; Rodzyanko, 1976; Iosifova, 1977; Zastrozhnov *et al.*, 1998; Grekov and Dolya, 1999). The peculiar, frequently armlike morphology of sand bodies stimulated us to analyze attentively all available data on their genesis. Most interesting in this respect is an area located westward of the southern termination of the Ergeni hills. At the beginning of 1960th, N.I. Luparev discovered in this area (near the village of Burukshan) a deep V-shaped incision (Rodzyanko, 1976) that was later investigated in detail by Kosarev and his colleagues (Kosarev *et al.*, 1965; Kosarev, 1971). They established that boreholes near the village of Divnoe (a near-wall zone of the incision) penetrated the Chokrakian inequigranular pebbly sands with dark limestone interlayers, which rest on the deeply eroded surface of the Ol'ga Formation (subdivision of the Maikop Horizon). According to Zhizhchenko, the limestone interlayers yield shells of *Acanthocardia centumpania* (Andr.), *A. impar* (Zhizh.), *Nuculana (Saccella) fragilis* (Chemn.), *Spiratella andrussovi* (Kittl), and some other, presumably early Chokrakian mollusks. Silty clay above this level yields small *Lutetia* shells suggesting the late Chokrakian age of host deposits. The overlying sequence includes the Karaganian beds with *Lutetia (Spaniodontella) gentiles* (Eichwald), and then the Konkian and lower to middle Sarmatian deposits. In addition, L.A. Kozyar distinguished here the palynological assemblage comparable with that characterizing the Chokrakian deposits of the Georgievskaya reference borehole (Kosarev *et al.*, 1965).

Later on, Rodzyanko (1976), referring to the opinion of A.I. Kopyl'tsov and K.G. Chemyakov who analyzed drilling results all along the incision up to the Ergeni foothills, concluded that the Chokrakian fossils are redeposited into the much younger (late Sarmatian–Maeotian) filling-in of the canyon, but we do not agree with this conclusion. First, neither Zhizhchenko, nor Kosarev mentioned features suggesting the redeposited state of fauna. Second, as is noted above, the Chokrakian interval was distinguished as the basal one in the complete Neogene sequence (from this level up to the middle Sarmatian inclusive), in which all organic remains and palynological spectra are successively stratified without signs of their mixing. This does not exclude, however, a possibility that there is a subsequent incision in the axial part of the paleocanyon. This incision was filled in with sediments during the terminal Miocene (after the mid-Sarmatian time), because Kosarev (1971) recovered rounded shells of Chokrakian and Karaganian mollusks from the base of the younger sedimentary filling-in.

Another area, where sands are widespread in the Chokrakian deposits, is located at the right side of the Terek River, within the sector of its latitudinal course. Southward of the Terek bench, in the northern slope of Terek-Caspian depression, a thick sequence (several hundreds of meters) composed of sand, clay, and silt with persistent land-sliding marks (Vassoevich, 1949) is traceable from Makhachkala to Groznyi (Zhizhchenko, 1940; Liverovskaya, 1951; Grossgeim, 1952). According to data of mineralogical analysis, sandy material was derived from the East European platform (Grossgeim and Gmid, 1951; Kholodov and Nedumov, 1981, 1996). In the northeast of the area under consideration, Grossgeim and Gmid distinguished a peculiar disthene–staurolite facies distinct from terrigenous material of the Caucasian origin. The avandelta of a river that delivered sandy material from the platform during the Chokrakian time was presumably located to the northeast of the Sulak River mouth, because sands deposited between this and Shura-Ozen' River are most coarse-grained, composed of subangular grains in contrast to better rounded material in southern and western areas (Grossgeim and Gmid, 1951). Facies of pebbly shore were distinguished in the Sulak River lower courses (Liverovskaya, 1951), and conglomerate outcrops are known in the Temirgoevskaya Ravine, at the Mt. Almalo, and near the Kaka Village northwestward of the Shura-Ozen' River (Zhizhchenko, 1940, p. 129). An impoverished assemblage of marine mollusks from these deposits that includes representatives of genera *Pholas, Mytilus* (Liverovskaya, 1951), *Donax*, and *Mactra* (Zhizhchenko, 1940) also suggests proximity to the former river mouth and accumulation of sediments under the desalinating influence of river runoff. Comparable environments were characteristic of the Karaganian time as well that is evident from found plant remains of *Sabal major* Heer, the palm species representing a component of plant groupings in marshy

settings near the coast (Grossgeim and Gladkova, 1950, with reference to Baikovskaya).

The considered sandy deposits, which are widespread in the Chokrakian sequences deposited in Cis-Caucasian region along the northern periphery of the basin, clearly show that an immense mass of terrigenous material drifted from the East European platform. The Paleo-Don river system was likely of prime significance in transportation of this material.

Zastrozhnov *et al.* (1998) demonstrated that the Paleo-Don river valley now filled up with deposits of the Yashkul' Formation, the lower horizons of which could be of the Chokrakian age according to palynological data, had several anabranches at the latitude of the Ovata Village, most of which were directed to the southeast and others to the south. Despite a partial erosion of the Yashkul' deposits, it is possible to suggest that a series of southeastern Paleo-Don anabranches corresponded to a delta with outer margin adjacent to avandelta located near the Sulak River mouth. This general paleogeographic situation is in accordance with that earlier outlined by Nedumov (1994). At the same time, the estuarine delta of the southern anabranch presumably joined a former canyon near the Divnoe Village. Iosifova (1977) who was of opinion that this canyon continued the Paleo-Don valley illustrated this idea in the paleogeographic map for the Miocene climatic optimum that is published in her recent work (Iosifova, 1999). From our viewpoint, the canyon that extends from the Gudilovskaya trough into the East Kuban' depression originated in the pre-Chokrakian time, most likely in the Kozachurian–Tarkhanian boundary interval, between the Sal'sk swell and Stavropol' arch, which were actively rising at that time. In opinion of Kosarev, the canyon developed during the Chokrakian time at the site of pre-Chokrakian strait between two parts of regressing sea basin. Many aspects of the canyon evolution and its morphology (depth and other parameters) are still problematic and should be better investigated in view of practical goals of prospecting geology.

For instance, it is difficult to bring our concept into line with occurrence of Tarkhanian deposits above the break of the eastern canyon slope near the Priyutnoe Village (Zastrozhnov *et al.*, 1998), and with their absence in the canyon itself. So, this is an intriguing area deserving a further study.

As is established in the central Cis-Caucasian region to the southwest of the canyon, where the Tarkhanian deposits are almost missing, the sand transport reached a steep northern slope of the West Kuban' depression (Mikerina and Pinchuk, 1997). Sand filled up channels in this slope or formed underwater fans within calcareous ooze near its base. Sandy material that accumulated along the southeastern wall of the trough was derived from the Greater Caucasus island (Grossgeim and Gmid, 1951; Bigun *et al.*, 1998). The western Cis-Caucasian region was partially supplied with sandy material via the ancient drainage system of the Severskii Donets River, alluvial deposits of which are traceable till the Taganrogskii Bay, and the Paleo-Mius River transported this material to the Black Sea depression (mapping results of A.V. Zaitsev, map sheet L-37-5, Shakhty, Yuzhgeologia). Within the Sea of Azov, like in the Dnieper and Azov–Kuban' depression, the continental fluvial deposits grade into shallow-water marine facies.

During the subsequent Karaganian time, when depressions quickly became filled up, the paleorelief flattened out, and the total shelf area shoaled so that transgressive events were perceptible along its marginal zone only. In the Chokrakian–Karaganian period, the West Kuban' and Terek–Caspian depressions approached configurations of the forthcoming foredeeps, and their shelf zone advanced to the south to be transformed in the eastern Caucasus into a coastal slope with mangroves (Kholodov *et al.*, 1980; Kholodov and Nedumov, 1996). In the western Caucasus, a deep peripheral trough bifurcating from the West Kuban' depression near Taman' received more distinct features than in the early Miocene. Its development along the uplift periphery was related to the progressing closure of the Caucasian marginal sea.

Until the recent time, the western part of the Turan plate was thought to be also occupied by the Chokrakian sea basin. The alternation of marine and continental sedimentation environments was suggested for the northern plate area (*Atlas...*, 1967). N.S. Solov'eva in Kunitskaya *et al.* (1969) argued for the former existence of two large sea gulfs interconnected via a wide strait that was situated in the west of the present-day Aral Sea.

Our revision and analysis of data on the Transcaspian region showed that sedimentation of the Chokrakian time developed here in three large gulfs. One, North Caspian, was localized in the northeast of the circum-Caspian depression, and another one corresponded to the vast North Ustyurt depression. The greater third gulf occupied areas of the South Mangyshlak, Assakeaudan, Verkhnii Uzboi, and Fore-Kopetdag depressions with synonymous smaller gulfs. It is likely that large gulfs did not communicate with each other, because areas separating them are lacking deposits with Chokrakian fossils. The Turan shelf, like that of the Cis-Caucasian region, was differentiated in the Chokrakian time. In addition to the North Ustyurt and Fore-Kopetdag depressions, which appeared earlier, the Chokrakian tectonic events resulted in downwarping of the northeastern circum-Caspian depression and present-day areas of the vast South Mangyshlak, Assakeaudan, and Verkhnii Uzboi depressions. Narrow elongated bays with lagoons in their rear parts extended from the last depressions to the Ilim–Balkuin and Sudoch'e depressions. This is evident, in particular, from occurrence of Chokrakian ostracodes in corre-

sponding deposits of the Balkuin depression (Rozyeva, 1976).

Individual gulfs were characterized by different types of sedimentation. The Chokrakian deposits in the eastern North Caspian and northern North Ustyurt depressions are represented by alternating marine and continental sediments, the former bearing polyhaline fauna. Concurrent deposits in the axial and southern areas of the North Ustyurt depression predominantly correspond to continental red beds intercalated with rare and thin interlayers of marine sediments. The North Ustyurt and North Caspian gulfs were separated by a wide land area that extended in the southwestern direction. The former existence of river system in this land is inferable from occurrence of sandy deltaic deposits in the northwestern Ustyurt, where they yield assemblages of mixed marine and fresh-water mollusks. At the same time, evaporite sedimentation was widespread in the South Mangyshlak, Assakeaudan, and Verkhnii Uzboi depressions (Eberzin, 1960; Chel'tsov *et al.*, 1967; *Neogenovaya sistema*, 1986a). It is clear that corresponding gulfs were isolated from the main Chokrakian basin by shoals, which hampered the water exchange. There could be some islands in the shoals, but it is impossible at present to establish their position and structure.

Northern gulfs of the Chokrakian basin functioned in the Trans-Caspian region during the Zyuk time and became exsiccated in the late Chokrakian. Some their remnants could exist, however, in the Bryk time as well, because there is a stratigraphic unconformity between the Chokrakian and overlying Miocene deposits, and consequently the upper Chokrakian sediments could be eroded. A late Chokrakian sea gulf presumably existed in the South Mangyshlak depression (Sharapov and Pleshcheev, 1979). In the marginal eastern part of the West Turkmen depression that represented a shallow shelf zone within the present-day area of western Kopetdag, sedimentation was in progress during the entire Chokrakian interval, because both the Zyuk and Bryk beds are distinguished here within the sandy–clayey sequence.

The Bol'shoi and Malyi Balkhan ranges likely represented two high islands situated in the shallow zone of the West Turkmen gulf. The northwestern periphery of the Bol'shoi Balkhan island corresponded to the accumulation zone of red, mostly rudaceous molasse beds with rare interlayers of marine sediments.

In distinction from western area, central to eastern region of the Kopetdag Mountains corresponded to a land area with hills in its most elevated zone. This rise was bounded in the northeast by a strike-slip fault zone that separated it from the Fore-Kopetdag depression. The eastern part of the latter and adjacent depression of southeastern Turkmenistan accumulated thick sequences of red, partly rudaceous molasse beds.

Vast plains of the Turan lowland, which were situated northward, represented areas of deposition of red terrigenous beds locally intercalated with gypsum-bearing interlayers of the lacustrine-proluvial, hyperhaline-lacustrine, and subaerial genesis. These deposits are preserved within a rather large area of the Kyzylkum desert, but northward, in the Circum-Aral and Circum-Caspian regions, they are almost completely eroded.

CONCLUSION

The Tarkhanian–Karaganian stage in the history of the Eastern Paratethys represents a single tectono-sedimentary cycle with characteristic regular variations of hydrological regime and biotic changes. These changes consisted in transition from the semiclosed Tarkhanian marine basin with the low rate of fauna endemism, first to the Early Chokrakian marine basin of deviating salinity, with a higher rate of endemism degree of fauna was higher, and then to the isolated Late Chokrakian and Karaganian basins with endemic biotas.

This study allowed us to depict, for the first time and rather completely, the morphology of sea basin that occupied a vast northern periphery of the Tethys along with the collision zone between the Eurasian and Afro-Arabian plates during the Chokrakian time. Currently accepted sedimentological concepts on the evolution and paleobathymetric zoning in marginal seas and the ecological-geographic analysis of biota evolution during the period under consideration were used to reveal a detailed picture of the basin dynamics. The established sedimentation environments, which are thoroughly described for the distinguished paleobathymetric zones, were, as is known, responsible for origin of petroleum deposits in the Chokrakian and Karaganian sequences of some regions (Terek–Sunzha interfluve and Dagestan piedmonts). The verified bathymetry, area and configuration of the basin area are also important for a deeper insight into tectonics of the region.

As is established, connections of the studied basin with seas of the Tethyan realm (ancient Mediterranean, Mesopotamia, and Indian Ocean), which resumed in the Tarkhanian, after the Kozachurian isolation, and were open during the early Chokrakian climatic optimum, resulted in transgression of the Eastern Paratethys due to the World Ocean level rise. Beginning since the late Chokrakian and during the Karaganian time, the basin isolated at that time expanded further because of changes in its water balance.

The intense tectonic movements of the early Chokrakian time considerably differentiated the shelf relief and were responsible for an extreme undulation of shoreline, especially in the Transcaspian region, where extended gulfs of regional depressions run as deep inland as several hundreds of kilometers. The peculiar bryozoan-algal bioherms, which have no analogues at present, appeared for the first time in the shoals considerably expanded at that time in the north of the Euxinian–Caspian basin.

Sedimentation in the northern shelf was under strong control of terrigenous influx from the East European plain, and the Paleo-Don river was of prime significance in the drainage system of the latter. The available data on deltaic deposits to the northeast of the Terek bench (Makhachkala area) and the peculiar structure of deposits filling up the paleocanyon in the central Cis-Caucasian region (Divnoe Village) were used to interpret anew the Paleo-Don evolution in its lower courses. By analogy with the present-day Volga and other large rivers, it is suggested that the Paleo-Don was divided here into anabranches, which met the sea far away from each other. This model, as we believe, settles the difference in existing views on position of delta of this greatest mid-Miocene fluvial artery of the East European plain.

The intense terrigenous sedimentation that quickly filled up the deep-water and shelf depressions of the Trans-Caucasian region during the Chokrakian–Karaganian time was a consequence of concurrent orogenic processes in the Anatolian–Iranian microcontinent. Depressions uncompensated by sedimentation remained only in the Kobystan–South Caspian part of the Mesozoic marginal sea of the Greater Caucasus.

Active tectonic movements of the initial middle Miocene time (Chokrakian–early Badenian–Langhian) resulted in restricted connections between the Chokrakian and Indo-Pacific basins, though without cessation of water exchange with the Central Iranian and, possibly, with Mesopotamian basins. Simultaneously, they resulted in separation of the Eastern and Western Paratethys. Shoaling in the central part of the Chokrakian basin gave rise to the appearance of the Caucasian archipelago. The mammalian Belomechetka fauna that originated in the early mid-Miocene time in the north of the Greater Caucasus island includes representatives of African and Eurasian species. This fossil assemblage implies an unstable bridge that existed in the Near East and southern Anatolia and made the mammalian fauna possible to interchange between the Afro-Arabian and Iranian plates, on the one hand, and Eurasia, on the other. The appearance of Caucasian land ceased transportation of terrigenous material from the north, the bulk mass of which accumulated in the Cis-Caucasian depressions, where the avalanche-like increase of sedimentation rate is recorded in the Chokrakian and, to some extent, in the Karaganian deposits.

Somewhat different sedimentation environments were characteristic of eastern (Transcaspian) basin areas. Thick evaporite beds deposited here suggest that middle and terminal stages of sedimentation in the period under consideration (especially in the Karaganian time) were the time of distinctly growing aridity of the climate.

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